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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

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THE GEOMORPHIC HISTORY OF THE UPPER WABASH VALLEY*

W. D. THORNBURY

ABSTRACT. Most of the area now drained by the upper Wabash Valley was in Tertiary time part of the Teays Valley drainage. Burial of the Teays Valley and its associated Tertiary topography and diversion of part of its drainage to the Wabash were initiated by the Kansan glaciation and completed by the Illinoian.

The upper Wabash River is superposed across a buried Tertiary topography that the river has hardly more than begun to exhume. Headward, from about Terre Haute, Indiana, the Wabash Valley consists of alternating wide and narrow stretches depending largely upon whether its course crosses a preglacial valley or a buried preglacial bedrock upland. The courses of the Wabash and Teays in Indiana roughly parallel each other but actually intersect at only two places.

The most conspicuous topographic features along the Wabash Valley are largely a consequence of its having been a major Wisconsin sluiceway. Most striking are: (1) minor scabland tracts, where the sluiceway cuts through buried bedrock uplands, (2) partially exhumed bioherms (klintar), where the valley crosses the belt of Silurian rocks, (3) numerous abandoned valley braids above the present valley floor, (4) gravel and bedrock terraces, and (5) dune and loess deposits in or adjacent to the sluiceway.

The surficial loess adjacent to the Wabash Valley is of Tazewell age, but buried loess that has been interpreted as of Farmdale age indicates that during this Wisconsin subage ice extended far enough into Indiana to send outwash down the Wabash Valley. It is questionable whether Iowan loess is present along the Wabash Valley. What has been previously called Iowan loess may be pro-Tazewell loess. No Cary loess has been mapped, but till and outwash of this stadial are present in the uppermost part of the Wabash drainage basin.

INTRODUCTION

The Wabash River is commonly thought of as a Hoosier stream, although its source is in Ohio about 12 miles east of the Indiana-Ohio boundary. Its lower course forms the boundary between Indiana and Illinois. Nearly three-fourths of the drainage basin of the Wabash and its tributaries is in Indiana. The portion here designated as the upper Wabash Valley is that part of the river from near Terre Haute, Indiana, headward to its source. There are two reasons for logically limiting the present discussion to this part of the Wabash Valley. The first is that the portion of the valley from Terre Haute southward to its junction with the Ohio has been previously studied and discussed (Fidlar, 1948), and the second is that the boundary here chosen essentially coincides with the southernmost extension of the Wisconsin ice sheet in Indiana.

The Wabash River, according to Fidlar, has a drainage basin of some 33,000 square miles, 300 of which are in Ohio, 8,600 in Illinois, and 24,000 in Indiana. The source of the Wabash is at an altitude of 870 feet, and its junction with the Ohio is at an altitude of about 320 feet. The gradient of the river ranges from about 3 feet per mile in the upper 100 miles to approximately 8 inches per mile in its lower course.

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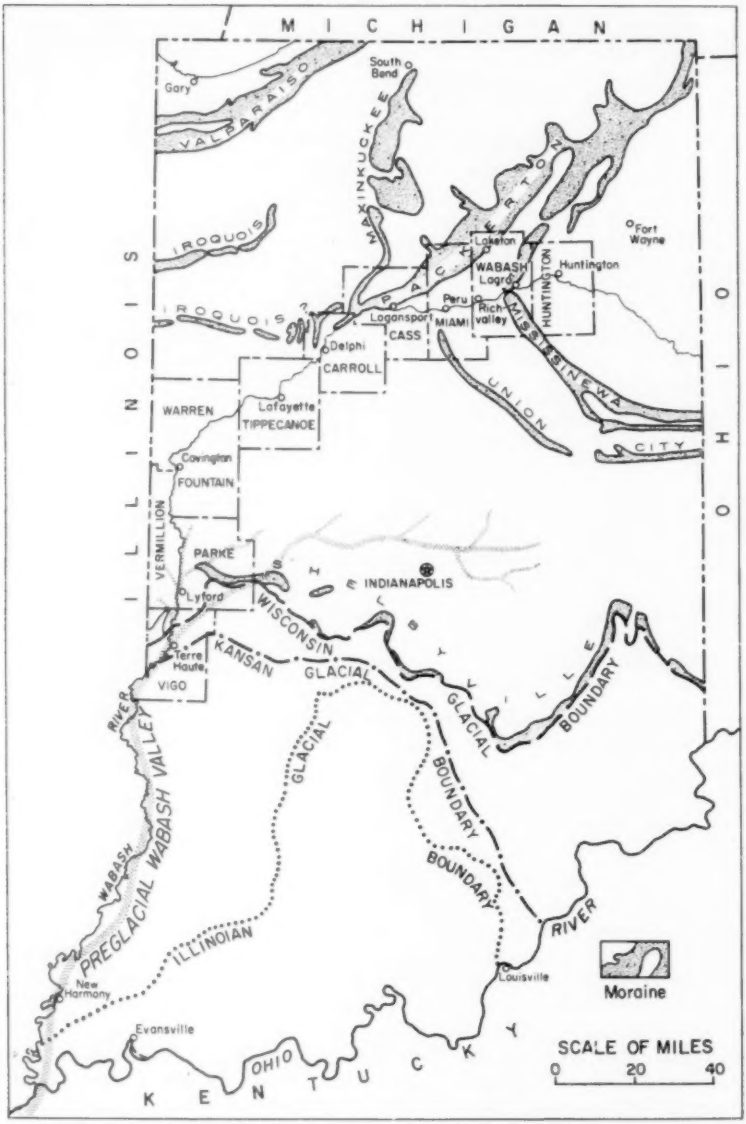

Fig. 1. Index map of Indiana showing towns, counties, glacial boundaries, and moraines to which reference is made in the text. Present and preglacial courses of Wabash River are also shown.

The Wabash River today is a considerably larger stream than it was in Tertiary time. Extensive drainage has been added to it as the result of filling of preglacial drainageways by Pleistocene glaciers. An explanation of the genesis of the present Wabash River is a primary aim of this paper. A second objective is a discussion of the topographic features along the Wabash Valley that are a result of the role played by the Wabash Valley as a major glacial sluiceway during the Wisconsin glacial age. To do this we must start back in the Tertiary.

TERTIARY DRAINAGE OF EAST-CENTRAL UNITED STATES

Probably as much work has been done on the preglacial drainage of Ohio, Indiana, and Illinois as on the preglacial drainage of any other comparable area in the United States, but differences of opinion as to its details still exist. The preglacial drainage of Illinois has been worked out more completely (Horberg, 1950) than has that of Indiana and Ohio, but, as the salient facts for the latter two states are reasonably well known, it is now possible to present the general picture fairly accurately.

There seems to be little question that what Tight (1903) designated as the Teays Valley in West Virginia and its continuation across Ohio (Ver Steeg, 1936), Indiana (Wayne, 1952), and Illinois (Horberg, 1945, 1950) was the major drainage line of east-central United States during Tertiary time. The Teays River headed in the Blue Ridge or possibly Piedmont of North Carolina and flowed northwestward across West Virginia and Ohio and then southwestward across Indiana into Illinois, where it joined the Mississippi River in Tazewell County, in west-central Illinois. It is probably more nearly correct to state that the preglacial Mississippi joined the Teays in central Illinois. The route of the Teays was rather remarkable in that its source was unusually near the Atlantic coast for a stream that had its terminus in the Gulf Embayment.

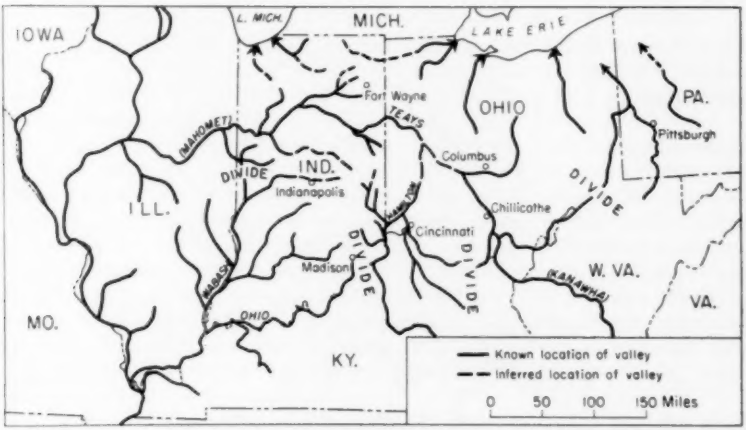

Fig. 2. Postulated preglacial valleys and divides in midwestern United States during late Tertiary time. (From Wayne, 1950 and based upon the ideas of numerous individuals).

The time of origin of this river system is uncertain. It may have been the major river of eastern United States throughout the Mesozoic as well as Cenozoic time. The part of its history which is decipherable is limited mainly to late Tertiary and early Pleistocene time. In considering the history of the Teays River we are concerned primarily with the obliteration of this drainage system by continental glaciation and the resulting diversion of part of the Teays drainage to the Wabash and Ohio Rivers.

The writer's ideas regarding the major drainage lines of east-central United States during Tertiary time are essentially similar to those shown in figure 2. The regional picture is fairly clear, but there are questions as to where some of the major stream divides were located. The position of the divide in northeastern Illinois between the drainage to the Gulf of Mexico and to the Gulf of St. Lawrence probably is essentially correct. There is difference of opinion as to whether the divide between the preglacial Ohio and Teays drainage was above or below Cincinnati.

PREGLACIAL DRAINAGE OF INDIANA

Studies in recent years have made possible a generalized reconstruction of the preglacial topography of Indiana. Wayne (1952) has presented a pic-

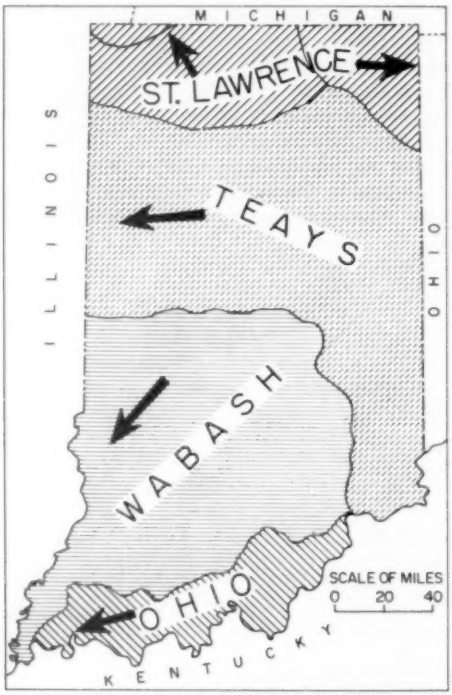

Fig. 3. Preglacial drainage basins of Indiana.

ture of the topography and drainage of that part of Indiana (approximately five-sixths) in which the preglacial topography is largely obscured beneath glacial drift. Figure 3 represents the writer's interpretation (based largely upon Wayne's work) of the major preglacial drainage basins of Indiana. The divide between the preglacial Wabash and Teays has been variously located by workers upon this problem. Fidlar (1948, fig. 1) placed it near Covington in Fountain County, Indiana. Thornbury (1948) thought that the divide might have been near the crest of the Knobstone escarpment where it extends across southern Tippecanoe County, northern Fountain County, and northern Warren County. Wayne's (1952) early studies of the bedrock topography seemed to indicate that the head of the preglacial Wabash was farther south in northern Parke County and southern Fountain County. More detailed work by Wayne (1956) later led him to conclude that the main branch of the preglacial Wabash was not the one that headed near the western boundary of Indiana but rather was a valley, named by him Montclair Valley, that headed in east-central Indiana, in Hancock County, on the blackslope of the Laughery escarpment, an escarpment formed on resistant Silurian limestone (Niagaran series).

From such knowledge a we now have it appears that the preglacial drainage of Indiana can be described in terms of five major drainage areas: drainage to the Ohio in southern Indiana, the Wabash drainage of southwest-central Indiana, the Teays drainage of southeastern and central Indiana, the drainage of northwestern Indiana into a lowland now occupied by Lake Michigan, and a small area in extreme northeastern Indiana that drained into a lowland now occupied by Lake Erie. Thus, in Tertiary time, distribution of runoff from Indiana was not significantly different from what it is today. The major difference was that the Teays River was the largest river in Indiana, whereas the Ohio and Wabash were much shorter and smaller streams than today.

TERTIARY GEOMORPHIC HISTORY OF INDIANA AND ADJACENT STATES

Little can be said about the geomorphic history of Indiana in pre-Tertiary time. Presumably Indiana has been above sea level since the close of the Paleozoic, but all evidence of Mesozoic erosion surfaces has been obliterated and even the early Tertiary history of the state is obscure.

The geomorphic histories of such unglaciated areas as south-central Indiana, Kentucky, Tennessee, the Appalachian plateau area of Ohio and West Virginia, and the Driftless Area of Wisconsin, Minnesota, and Iowa, all areas that may be presumed to have had geomorphic histories similar to that of central and northern Indiana, are fairly well known. Opinions as to how many cyclical erosion surfaces should be recognized in the Mississippi-Ohio Valley region vary from none (Martin, 1916) to as many as five (Hershey, 1896). Most workers have recognized evidence of one or two major cycles that attained the peneplain or near peneplain stage and some have seen what they considered evidence for one or more subcycles now represented by strath terraces.

In the Driftless Area of Wisconsin two peneplains have been described (Trowbridge, 1921), the higher and older Dodgeville and the lower and younger Lancaster. Trowbridge (1954), however, recently concluded that there is

really only one peneplain, the Lancaster, present. This peneplain surface lies 300 to 600 feet above present valley flats and 625 to 725 feet above buried bedrock valley floors. Presumably, the cycle during which the Lancaster peneplain was formed corresponds with that during which the Highland Rim peneplain of Tennessee (Hayes, 1899) and the Lexington peneplain of Kentucky (Campbell, 1898) were formed. This cycle of peneplanation in the eastern interior United States apparently corresponds in a general way with the Harrisburg cycle of the Appalachians. The Dodgeville peneplain of the Driftless Area, if it is really a higher and older surface than the Lancaster, would thus seem to correspond to the Schooley cycle farther east. Little evidence has been found in the area south of the glaciated region of an erosion surface older than that produced during the Highland Rim-Lexington cycle. Some workers have considered the tops of hills that project above the level of the Highland Rim surface to be remnants of a higher and older erosion surface, but evidence for a pre-Highland Rim-Lexington peneplain in Indiana, Kentucky, and Tennessee is rather meager.

Malott (1922) recognized in southern Indiana an early Tertiary peneplain that he correlated with the Highland Rim and Lexington peneplains. Remnants of this erosion surface are found at altitudes of 900 to 1,000 feet in southeastern Indiana. Malott considered this peneplain to be of probable Eocene age, but it is now more commonly considered to be of Miocene or Pliocene age. Malott also described in southern Indiana an erosion surface below the Lexington-Highland Rim peneplain that he called a late Tertiary peneplain. The cycle that produced this erosion surface destroyed much of the so-called early Tertiary peneplain in western Indiana. The late Tertiary peneplain of Malott is probably better described as a strath terrace than as a peneplain.

Horberg (1946, 1950), from his studies of the bedrock topography of Illinois, thought that four buried erosional surfaces are recognizable beneath the glacial deposits of Illinois. These he designated as the Dodgeville, Lancaster, and Central Illinois peneplains and the Havana strath. The Lancaster peneplain was correlated with the Calhoun peneplain (Rubey, 1952) of southwestern Illinois and the Ozark peneplain of Missouri (Marbut, 1896; Hershey, 1901). Remnants of the Dodgeville peneplain are found only in a small area in northwestern Illinois and are at altitudes between 1,000 and 1,150 feet above sea level. The more extensive Lancaster surface lies about 150 feet below the Dodgeville surface. The most extensive erosion surface in Illinois is the Central Illinois peneplain. It is found consistently at altitudes between 600 and 650 feet. It is most widespread on upland tracts to the south and north of the Mahomet (Teays) valley of central Illinois. Horberg (1946, 1950) thought that there is evidence of strath terraces along the preglacial Mahomet, Mississippi, Kaskaskia, and Wabash valleys. These terrace tracts are found at altitudes ranging from 550 feet at the north to 450 feet or less in southern Illinois and were interpreted as remnants of a planation surface developed before the period of valley entrenchment that resulted in the so-called Deep Stage. Horberg named this erosion surface the Havana strath. Entrenched bedrock valleys 100 feet or more below the Havana strath represent the last period of valley erosion prior to burial of the preglacial topography beneath glacial deposits.

There has been difference of opinion as to when the Deep Stage was cut but Horberg believed that the valley deepening in Illinois took place prior to Nebraskan glaciation. Ver Steeg (1936) suggested that cutting of the Deep Stage started after disruption of the Mahomet-Teays drainage and continued until the advent of Illinoian glaciation. Ireland (1946) considered the valley trenching to have been additive and thought that degradation occurred during the early stages of each glacial age as a result of lowering of sea level and consequent increase in stream gradients. He thought that degradation occured during early phases of the Nebraskan, Kansan, and Illinoian glaciations and was terminated when the extensive Illinoian glaciation disrupted many major drainage systems. Evidence cited below for extensive Kansan glaciation east of the Mississippi River makes it seem unlikely that major Tertiary drainage lines persisted in the glaciated area as late as Illinoian time.

PLEISTOCENE HISTORY OF AREA

Nebraskan Glaciation.—Whether the Nebraskan glaciation affected Indiana is still questionable. No unequivocal evidence of Nebraskan deposits has been recognized in Indiana. Thwaites (1946, pl. 3), largely upon the basis of glacial erratics described by Leverett (1929) in northeastern Kentucky, postulated a Nebraskan ice lobe that extended across southeastern Indiana and southwestern Ohio into Kentucky. There is no convincing evidence of the Nebraskan age of these boulders. It can only be said with assurance that they are probably pre-Illinoian in age, and, as will be shown later, there is a greater probability that they are Kansan than Nebraskan.

Eveland (1952) assigned a Nebraskan age to a large block of gumbotil embedded in Kansan till in a strip mine near Danville, Illinois, but personal examination of the exposure left the writer uncertain as to the correctness of this interpretation. Horberg (1953) described what he considered to be buried Nebraskan drift in northeastern Illinois and also in central-western Illinois (1956) and suggested the possible presence of Nebraskan deposits in the bedrock Mahomet and Danvers Valleys of central Illinois. If these interpretations are correct, the Nebraskan ice sheet may well have extended into northern Indiana and may have crossed the Teays Valley, but as yet no uncontrovertible evidence of this ice invasion has been recognized in Indiana.

Kansan Glaciation.—The evidence for Kansan glaciation in Indiana is much stronger than that for Nebraskan. MacClintock (1929) described several exposures of pre-Illinoian till that are in Illinois and slightly southwest of Terre Haute, Indiana. He was uncertain as to whether the tills are Nebraskan or Kansan in age, but in view of our present knowledge of the extent of those two glaciations in the Midwest, a Kansan age seems more likely. Wayne (1954, 1955, 1958) has described several exposures of what he considered to be Kansan till beneath calcareous Illinoian till in Indiana. There are several exposures of Kansan till in the area east and northeast of Terre Haute. These fit in very nicely with the occurrences of Kansan till in Illinois. Beneath till at one exposure is a calcareous and fossiliferous loess that contains several species of gastropods suggestive of a Kansan age. The loess is calcareous throughout and is probably a pro-Kansan deposit closely related in age to the till above it. Ex-

posures of Kansan till in southeastern Indiana are so situated as to suggest that a lobe of Kansan ice extended into southeastern Indiana as far as, if not beyond, the Ohio River (see fig. 1). The rapidly accumulating evidence for extensive Kansan glaciation in Indiana, particularly in southeastern Indiana, suggests the probability that the erratics in Kentucky that Leverett (1929) described and Thwaites (1946) interpreted as Nebraskan in age are more likely of Kansan age.

Illinoian Glaciation.—The extent of Illinoian glaciation in Indiana is rather well known, (see fig. 1) as a result of the mapping of the glacial boundary in Indiana by numerous workers. Until recently it was assumed that the Illinoian ice extended farther into Indiana than did any other ice sheet, but the recognition of Kansan drift in southern Indiana makes this assumption one of uncertain validity. It has already been established that at one locality in south-central Indiana the Kansan ice reached slightly farther south than did the Illinoian and this may be true in other parts of the state. This is particularly likely in southeastern Indiana where, as stated above, it appears probable that the Kansan glaciation reached into Kentucky.

Wisconsin Glaciation.—The boundary of Wisconsin glaciation in Indiana is readily recognizable. Presumably the surficial Wisconsin tills of Indiana

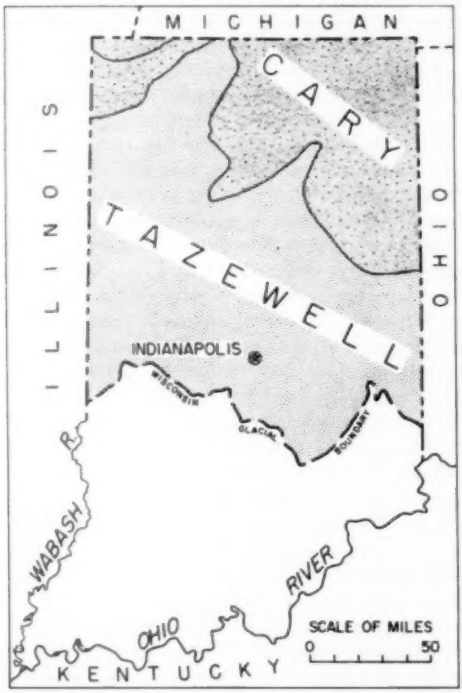

Fig. 4. Areas of Tazewell and Cary tills in Indiana according to the writer's concept.

belong to the Tazewell and Cary substages (fig. 4). The possibility exists, however, that some of the till that has been considered Tazewell in age may actually be of Farmdale or Iowan age. Farmdale loess has been recognized at several localities in the Wabash Valley area and loess that was interpreted as Iowan loess has been described (Thornbury, 1937) north of the outer Wisconsin border. The writer now believes that at least some of the loess that he earlier interpreted as Iowan is more logically considered to be pro-Tazewell loess, although there may be relatively little difference in age between an Iowan loess and a pro-Tazewell loess in Indiana. There seems to be no question about the presence of Farmdale loess in the lower Wabash Valley and this must mean that the Farmdale glaciation reached far enough into Indiana to send outwash down the Wabash Valley. Ray (1957) has recently described what he considered to be Farmdale loess in the vicinity of Louisville, Kentucky, and if this interpretation is correct, the possibility should be recognized that Farmdale ice somewhere sent outwash down the Ohio Valley, for the Farmdale loess at Louisville seems to be too far away to have been derived from outwash in the Wabash Valley. Furthermore, the Farmdale loess in the Louisville area seems too thick to have been derived from the Wabash Valley. It is 5± feet thick, according to Ray, whereas the greatest known thickness of Farmdale loess in the Wabash Valley is only 4 feet.

On the glacial map of North America (Flint and others, 1945), except for the area back of the Valparaiso moraine in northwestern Indiana, which is mapped as Cary in age, the Wisconsin glacial deposits of Indiana are shown as undifferentiated. Wayne and Thornbury (1951) concluded that the Mississinewa moraine marked the outer limit of Cary glaciation by the Erie lobe. The writer now believes that the Union City moraine (see fig. 1) more likely is the outer moraine of the Cary glaciation in the area covered by the Erie lobe. This moraine marks a definite change in till petrology, the till back of it having a higher clay content than does the Tazewell till beyond it. Radiocarbon dating (Suess, 1954) of some gyttja from a marl pit at the southeast edge of the Packerton moraine, near the village of Laketon, in Wabash County (see fig. 1) gave an age of 13,140 ± 400 years. This indicates that the Packerton moraine of the Saginaw lobe is early Cary in age. The writer, at present, considers the Packerton and Maxinkuckee moraines to be the outer Cary moraines of the Saginaw lobe, but this conclusion is very tentative and is based primarily upon the topographic change which takes place at these moraines rather than upon any careful study of till petrologies. Lakes and closed basins are common within these two moraines but are scarce outside them. Actually the Cary drift of the Saginaw lobe has been so little studied in Indiana that the location of the outer Cary deposits is questionable. Horberg (1955) mapped Cary drift of the Saginaw lobe across Indiana a short ways into Illinois and considered the Iroquois moraine (see fig. 1) to be the end moraine of this substage. Other Illinois geologists share this viewpoint. Only by detailed field study and radiocarbon dating of the deposits in this part of Indiana can we ultimately determine where the boundary for the Cary substage is in the area of the Saginaw lobe.

EFFECTS OF GLACIATION UPON PREGLACIAL DRAINAGE

The major effect of the several glaciations of Indiana was obliteration to a large degree of preglacial drainage lines in northern and central Indiana and superposition of new stream courses across the buried bedrock topography. The Teays Valley was completely buried and the Wabash Valley became the major drainageway of central Indiana. Most likely it was the Kansan glaciation that largely caused the obliteration of the Teays as a major through drainageway. Nebraskan ice may have extended across the Teays Valley in Indiana and Illinois, but it seems probable that even if it did the effect was not a closing of the Teays as a regional drainageway.

Horberg (1945, 1953) attempted a reconstruction of the topography of central and northeastern Illinois during Sangamon time, and he concluded that by then the Mahomet (Teays) Valley had been largely obliterated as a drainage line in Illinois. Sufficient data are lacking at present to make possible a reconstruction of the Sangamon surface in Indiana, but in view of the newly discovered evidence that Kansan ice extended far south of the Teays Valley in Indiana, it sems most likely that it was this ice invasion that initiated burial of the Teays Valley. It is therefore postulated that as a result of Kansan

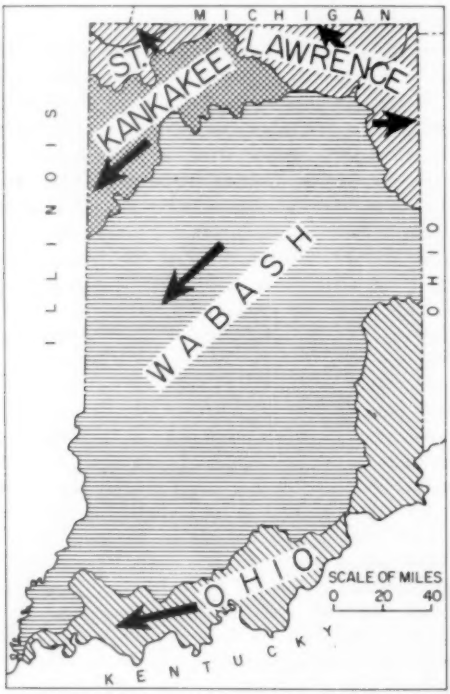

Fig. 5. Present drainage basins of Indiana.

glaciation the following modifications of preglacial drainage were effected: (1) upper Teays drainage as far north as southern Ohio and southeastern Indiana was diverted to the south and joined with the preglacial Ohio whose headwaters were in southeastern Indiana or southwestern Ohio, and (2) most of the Teays drainage area in central Indiana was added to the preglacial Wabash (fig. 5). This interpretation is not meant to imply that the Teays Valley in Indiana was completely buried by Kansan glaciation and abandoned, for it is probable that stretches of the Teays Valley continued to function as local drainage lines. Illinoian glaciation further obscured and closed portions of the Teays Valley, but discontinuous segments still served as drainage lines as late as Wisconsin time, as is indicated by the extensive outwash gravels beneath Wisconsin till in portions of the Teays Valley in Wabash, Miami and Tippecanoe Counties.

GEOMORPHIC FEATURES OF WABASH VALLEY

Relation of Courses of Teays and Wabash

The present route of the upper Wabash came into existence largely following deglaciation during the Tazewell and Cary subages of the Wisconsin age. Although the upper Wabash follows in a rough way the same general course as did the preglacial Teays, at only two localities (fig. 6) do the two river courses

Fig. 6. Comparison of courses of Teays and Wabash Valleys across Indiana.

coincide. One of these is just east of the hamlet of Rich Valley in western Wabash County and the other is near Lafayette in Tippecanoe County. At both places, there is notable widening where the Wabash Valley intersects the buried Teays because the drift-filled Teays Valley is more erosible than the stretches to either side where the present course of the river extends through preglacial bedrock uplands.

Partial Exhumation of Tertiary Topography.—As a result of the superposition of the Wabash Valley on glacial drift and its lack of coincidence with the course of the Teays, the Wabash Valley consists of more or less alternating stretches of wide and narrow valley as it crosses preglacial valleys and divides. At the intersection of the Wabash and Teays Valleys east of the village of Rich Valley, referred to above, (fig. 7) there is an expansion of valley width

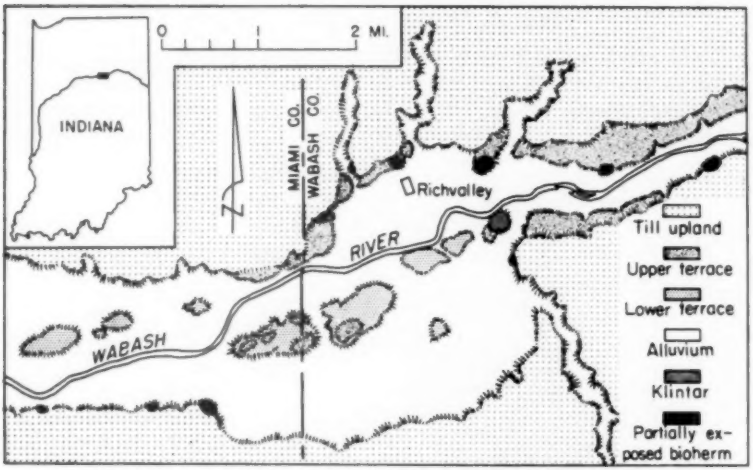

Fig. 7. Widening of Wabash Valley at its intersection with Teays Valley.

from about half a mile where the valley is cut through Silurian limestone and bioherms to nearly 3 miles where it intersects the Teays Valley. An almost equally striking expansion in valley width related to partial exhumation of the Teays Valley takes place in the Lafayette area. The great width of the Wabash Valley here is in part a result of a major tributary valley, Anderson Valley (Wayne, 1956), joining the Teays at this point. These preglacial valleys were filled with Wisconsin outwash, and thus the widest gravel terrace tract to be found along the entire length of the upper Wabash was formed. The filled valley is nearly 5 miles wide, but the present alluvial strip is hardly 1 mile wide.

The width of the Wabash Valley in those stretches where it is cut into bedrock reflects to a large degree the ease with which the rocks into which it is being cut are eroded. Where the valley crosses Silurian rocks, it is notably narrower through the outcrop area of the cherty Liston Creek limestone than

across the Mississinewa shale. In the belt of Devonian rocks the valley is much narrower where it crosses limestones than across the New Albany shale. The Pennsylvanian sandstone in the Pottsville series causes a marked narrowing of the valley near the west margin of Tippecanoe County.

Bioherms and Klintar

Possibly the most interesting bedrock features to be seen along the Wabash Valley are the many Silurian bioherms that have been partially resurrected. These features, often referred to as reefs, have cores of massive dolomite that are more resistant to erosion than the associated interreef shales and limestones. The bioherms consequently have topographic expression as bold bluffs along the valley sides or occur along the valley floor as hills or low mounds known as *klintar* (singular, *klint*) (Shrock, 1928). A klint located just south of the Wabash River 1 mile southeast of the village or Rich Valley

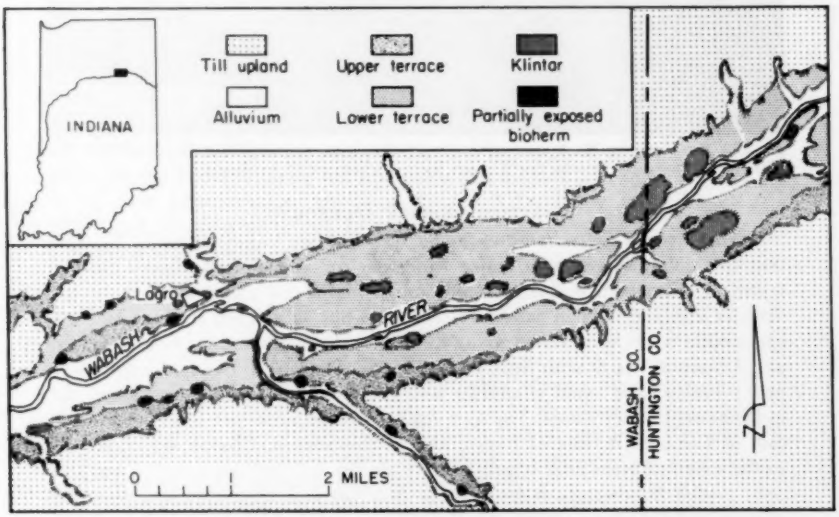

Fig. 8. Bioherms and klintar in Wabash Valley east of Lagro, Indiana.

and about 2 miles east of the west county line of Wabash County rises abruptly about 80 to 90 feet above the valley floor. Probably the most famous klint is Hanging Rock on the south bank of the Wabash River about 1 mile southeast of Lagro in Wabash County (fig. 9). During its exhumation by the Wabash River about half of the bioherm has been eroded, and the remainder stands about 75 feet above the river level. About 90 bioherms have been mapped in the area of outcrop of Silurian rocks in Miami, Wabash, and Huntington Counties. These bioherms commonly are reflected in the topography as klintar, by steepening of the valley wall where they crop out, or as water falls or rapids in tributaries of the Wabash.

Fig. 9. Hanging Rock, east of Lagro, Indiana, an exhumed Silurian bioherm. The bioherm has been cut into by the Wabash River. (Photo by C. A. Malott).

Miniature Scabland Topography

Many topographic features along the Wabash Valley attest to the fact that this valley served as a major sluiceway for glacial melt waters. Several tracts of miniature scabland topography, similar in many respects to the famous scabland area in eastern Washington but on a much reduced and less spectacular scale, are present along the valley. The scabland consists of water-swept bedrock tracts with little or no glacial outwash on them, anastomosing valley braids, and small islandlike bedrock remnants that project above the torrent-swept valley floors. The four main scabland areas are found at Delphi, east of Logansport, east of Lagro, and at Huntington. Klintar are numerous in the scabland tract near Lagro (fig. 8). In the scabland area east of Logansport glacial floodwaters flowed across Silurian and Devonian limestones in numerous braided channels. In the abandoned valley braids many small patches of bedrock crop out through a thin veneer of outwash, and numerous small mounds of Devonian limestone not quite reduced by the glacial floodwaters stand a few feet above the floor of the sluiceway. Pulpit Rock, on the south side of the Wabash Valley about 5 miles east of Logansport, is one of the better-known examples of these small rock nubbins. It consists of a mass of Kenneth (Devonian) limestone resting upon Kokomo limestone of Silurian age.

VALLEY TRAIN DEPOSITS

The importance of the Wabash Valley as a glacial sluiceway is shown by the amount of glacial outwash found along it. Gravel fill is found chiefly in those areas where the valley crosses or follows preglacial or interglacial val-

leys. The thickness of the fill varies greatly and exceeds 200 feet in many places. A well in the valley near Peru, in Miami County, penetrated 237 feet of gravel and mud, and several wells in the gravel terrace upon which Purdue University is located at West Lafayette, in Tippecanoe County, go through 200 to 210 feet of sand and gravel before reaching bedrock.

Abandoned Valley Braids

Many large abandoned valley braids that no longer carry water except at times of extremely high floods, such as those of 1913 and 1937, further attest to the great volume of glacial melt water than once flowed down the Wabash sluiceway. These abandoned valley braids enclose numerous island-like tracts of both bedrock and outwash gravels that stand as terraces in the sluiceway.

TERRACES

Terraces are probably the most conspicuous features found along the Wabash sluiceway. Bedrock terraces characterize the scabland areas but elsewhere the terraces are alluvial. They may occur as narrow strips abutting against the valley walls, as expansive areas several miles wide, or as detached tracts between valley braids in the sluiceway. Two terrace levels persist throughout the upper Wabash Valley, although at places there are suggestions of additional terraces, but it appears that these represent portions of one or the other of the two persistent terrace levels that have been reduced somewhat below their representative altitudes. Wier and Friedman (1955) described what they considered to be three terraces along a 25-mile stretch of the Wabash Valley extending above and below Terre Haute, Indiana. The topographic maps of the area suggest the presence of three terrace levels in this area but there is some question as to whether these three terrace levels persist throughout the upper Wabash Valley. Wier and Friedman's interpretation of the three terraces was that the highest terrace represents the upper limit of valley fill with Wisconsin outwash; the middle terrace they considered an erosion level developed along the valley prior to the discharge of overflow waters from Lake Maumee down the Wabash Valley; the lowest terrace was considered to represent a planation surface cut by the overflow waters from Lake Maumee or the so-called Maumee torrent.

Fidlar (1948) described two terraces in the lower Wabash sluiceway that he designated as the Shelbyville and Maumee terraces. The Shelbyville terrace is about 75 feet above the present floodplain in the area immediately south of the Wisconsin glacial border, and the Maumee terrace lies about 35 feet below it and 20 feet above the floodplain. The interval between the two terraces gradually decreases downstream, and near the junction of the Wabash with the Ohio the two terraces are only 20 feet apart. Fidlar interpreted the Shelbyville level as representing the upper level of valley filling with Wisconsin outwash and the Maumee level as a planation surface cut by overflow waters from Lake Maumee (the Maumee torrent) during the time that this ice-marginal lake drained through the Fort Wayne outlet to the Wabash Valley.

Terraces comparable in position to those in the lower Wabash Valley continue up the valley roughly to the neighborhood of Huntington (figs. 7

and 8) where the spillway of the Fort Wayne outlet joins the present Wabash Valley. Wayne and Thornbury (1951) described two terraces in Wabash County, the upper of which they called the Mississinewa terrace and the lower the Maumee. Both were considered to be of Cary age. The Mississinewa terrace was thought to be comparable in origin to the Shelbyville terrace of Fidlar but to be younger in age. They concurred with Fidlar's explanation of the Maumee terrace.

Part of the gravel fill in the upper Wabash Valley may be pre-Wisconsin in age, but thus far such outwash has not been identified. Deposition of Wisconsin outwash probably began as early as Farmdale time, for, as stated above, Farmdale loess has been identified in the lower Wabash Valley as well as at several places within the outer Wisconsin drift area. Whether the Iowan glaciation extended into Indiana is uncertain.

The tendency in recent years to divide the Wisconsin into numerous substages has greatly stimulated study of Wisconsin stratigraphy, but may have lead to overemphasis of the idea of prolonged breaks between the substages. It seems to the writer that the field evidence indicates that there were no long intervals between deposition of the successive till sheets. There appear to have been, in Indiana, three main stadials during the subage that has commonly been designated as Tazewell. Numerous well records commonly show three tills separated by outwash, and at several places within the area of Tazewell drift there are exposures of three Wisconsin tills. At no place is there a suggestion of a soil or weathered zone on any of the tills except the surficial one. Which particular advances of the Wisconsin ice sheet deposited these tills is as yet uncertain. Probably they are to be correlated with the ice advances that built the Shelbyville, Champaign and Bloomington moraines, but an earlier glaciation, such as the Iowan or Farmdale, is not to be ruled out.

At numerous localities in central Indiana a thin fossiliferous loess is found between what is conventionally thought to be the Shelbyville and Champain tills. Radiocarbon dates, as yet unpublished, indicate that there may be as much as 800 to 1000 years difference in the time of deposition of the two tills. Ages around 20,000 years suggest that the tills are of Tazewell age rather than older.

There may be some question, however, as to whether what have been considered the Shelbyville and Champaign stadials in Illinois and Indiana are exactly comparable. The possibility exists that what Leverett mapped as the Champaign moraine in Indiana is not exactly the same thing as the Champaign moraine in Illinois. Three lines of evidence suggest that this stadial may actually be as significant as some of the advances that elsewhere in the midwest have been given subage rank. The radiocarbon dates indicate at least a moderate withdrawal of the ice as does the presence of a fossiliferous loess at numerous places. Probably the strongest reason for suggesting that the so-called Champaign stadial in Indiana was just as significant as the earlier Shelbyville and later Cary advances is the soil pattern on the Wisconsin tills in the area covered by the Erie lobe. A glacial geologist can hardly look at the new *Map of Indiana Soils* published in 1957 by the Agronomy Department

of Purdue University without seeing that there are three distinct soil belts in the area glaciated by the Erie lobe. The soils of the Fincastle-Russell catena occur in the area from the Shelbyville moraine to the edge of the so-called Champaign moraine; the soils of the Miami-Russell catena lie in the region between the Champaign moraine and the Union City moraine; and the soils of the St. Clair-Nappanee catena occur in the area commonly believed to be the area of Cary drift.

There was more or less continuous outwash going down the Wabash Valley during the Tazewell subage, but there is reason to believe that during the Tazewell-Cary (Brady) interval[1] the ice withdrew a considerable distance to the northward so that for a time no outwash went down the Wabash sluiceway. The basis for this conclusion lies more in the difference in the Tazewell and Cary till lithologies than in the presence of a widespread soil profile upon buried Tazewell drift. Cary till in northeastern Indiana is notably higher in clay content than is the Tazewell till. A logical explanation for the larger amount of clay in the Cary till is that the clay was obtained from lacustrine deposits laid down in ice-marginal lakes in northeastern Indiana and northwestern Ohio while the ice front was many miles to the north.

Outwash was again carried down the Wabash sluiceway during the Cary readvance. After the Cary ice had retreated to the northeast of the Fort Wayne moraine, Lake Maumee came into existence and for a time it poured its waters, the Maumee torrent, through the Fort Wayne outlet down a broad spillway to the Wabash River at Huntington, Indiana.

Here, perhaps, a distinction should be drawn between the Wabash Valley and the Wabash sluiceway. The Wabash Valley heads in Ohio about 12 miles east of the Ohio-Indiana state boundary, whereas the Wabash sluiceway heads in Indiana at the Fort Wayne outlet of Lake Maumee. The two are one and same from Huntington downstream but the 25-mile trench from Fort Wayne to Huntington, followed through most of its length by U. S. Highway 24, is not a part of the Wabash Valley but was the spillway for Lake Maumee (fig. 10) in its earliest phase. It was not cut by the underfit Little River or Little Wabash that now flows through part of it, but rather by a great torrent of glacial melt water overflowing from Lake Maumee. This water was relatively clear water because Lake Maumee had acted as a settling basin for any outwash that came into it. Hence, this torrent was able to trench the fill that had been deposited in the Wabash Valley during Tazewell and early Cary time and thereby develop an erosional surface about 25 to 30 feet below the top of this fill. In this way the terrace was formed that Fidlar (1948) called Shelbyville and Wayne and Thornbury (1951) called the Mississinewa. Later opening of a lower outlet for Lake Maumee than that at Fort Wayne caused the Maumee torrent to run dry. Development in late glacial and Recent time of a floodplain below the Maumee erosion surface has left portions of it as terrace remnants

[1] There may be some doubt as to whether the term Brady commonly applied to this interval is entirely appropriate in Indiana, as there remains some question as to where the Brady soil belongs in the Wisconsin stratigraphic sequence. Detailed search for Brady soil horizons in northeastern Indiana has not been particularly fruitful. At a few places infant soil profiles have been found, mainly on gravels, but they are not very convincing that in Indiana there was a long break between Tazewell and Cary glaciations.

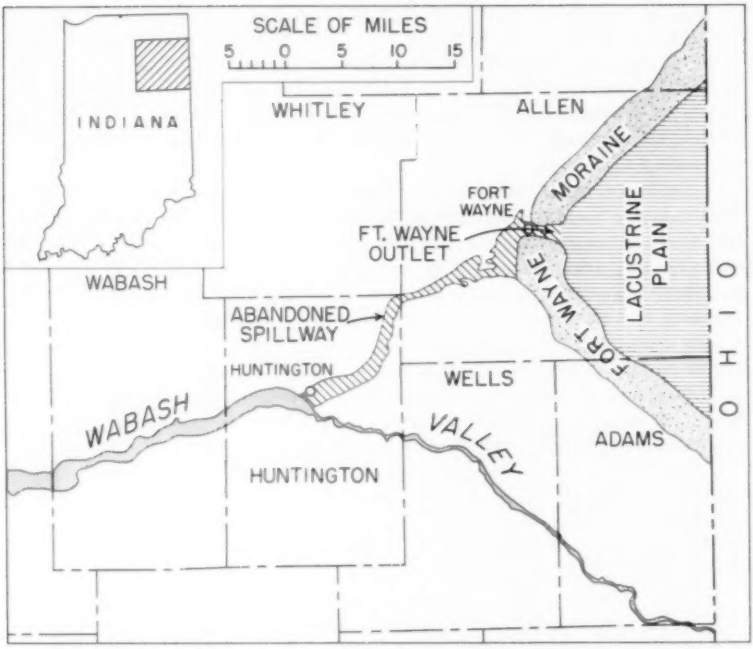

Fig. 10. Map showing relations of Lake Maumee and Fort Wayne outlet to Wabash sluiceway.

about 20 to 25 feet below the Mississinewa terrace and 15 to 20 feet above the floodplain of the river. It was largely the Maumee torrent that produced the scabland areas described above.

A different interpretation of the terraces would be to consider them climatic terraces. According to this viewpoint, glacial time was marked by valley aggradation and interglacial (or in this case intraglacial or interstadial) time by valley trenching. The upper terrace would represent the level of valley filling during the Tazewell subage. Presumably during the Brady interval this fill was trenched and partially removed. During the Cary glaciation, outwash again was deposited down the sluiceway, but this fill did not reach as high as the previous Tazewell fill. Dissection of the Cary valley train has left part of it as the Maumee terrace. No Mankato outwash is present because the outer limits of the Mankato ice sheet lay considerably north of the Wabash Valley.

It is difficult to say which of these interpretations is correct. The first one was advocated by Malott (1922), Fidlar (1948), and Wayne and Thornbury (1951). At the present time the writer is not completely convinced that it is the correct one, but he has been unable to find convincing evidence for the latter interpretation. Examination of many exposures of outwash sands and

gravels along the Wabash sluiceway has failed to produce evidence that there are two gravel fills in the valley.

If there has been trenching of the Tazewell fill during the Brady interval and later deposition of Cary outwash within the trenched gravels, it would appear that somewhere there should be evidence of unconformable relationships of the two gravel deposits. Such evidence has not been found. The fact that there is only one terrace along Eel River, an important tributary that enters the Wabash from the northeast at Logansport, lends support to the idea that the Maumee terrace is genetically related to the Maumee torrent. No terrace corresponding to the Maumee terrace exists in Eel Valley because Eel Valley had no episode in its history comparable to the Maumee torrent in the Wabash sluiceway.

Dunes

Windblown deposits are characteristic features along most glacial sluiceways, and the Wabash sluiceway is no exception. Extensive dune and loess deposits characterize the lower Wabash Valley (Fidlar, 1948); and, although not as extensive in the upper Wabash Valley, they are present. Dunal areas particularly are local in extent, but small dune areas exist in Parke, Fountain, Tippecanoe, Carroll, Cass, and Miami Counties. Small dunal tracts are found in most places where the sluiceway is wider than normal, but as the width of the upper Wabash sluiceway is not as great as that of the lower Wabash, sand dunes are not particularly conspicuous in size or extent.

The loess belt which lies to the east of the lower Wabash Valley continues northward into the area of Wisconsin drift; and, although the belt of thick loess becomes narrower, loess forms a more or less continuous blanket over the uplands on both sides of the sluiceway as far north as Warren and Fountain Counties.

CONCLUSIONS

1. The Wabash River was a much smaller stream in preglacial time than at present, but its drainage area has been notably increased as a result of the acquisition of part of the Tertiary Teays drainage after burial of the Teays Valley by the Kansan and Illinoian ice sheets.
2. Located as it is in an area of multiple continental glaciation, the Wabash Valley has served repeatedly as an important glacial sluiceway, being equaled or exceeded in importance only by the Illinois, Mississippi, Ohio, and Missouri Valleys.
3. The Wabash Valley is superposed across a buried Tertiary topography with little regard to the position of the preglacial valleys and uplands and the stream in the valley is now engaged in exhuming portions of the preglacial topography.
4. The striking topographic features along the Wabash Valley, such as abandoned valley braids, gravel terraces, scabland tracts, klintar, and sand dunes and associated upland loess deposits, are largely a consequence of the role that the valley played during Wisconsin time as a major glacial sluiceway.

5. Because of the importance of the Wabash Valley in the glacial history of central United States, its valley and adjacent till uplands provide an excellent area in which to study the effects of multiple continental glaciation and to work out the glacial stratigraphy of the Pleistocene.

ACKNOWLEDGMENTS

The present paper is an outgrowth of several years of areal mapping on a county basis in the upper Wabash Valley, the results of which have been or will be published by the Indiana Geological Survey. The writer wishes to thank Dr. William J. Wayne, Jacob Hamilton, Harold Deane, Bert Fell, and Paul Raymond, who were field assistants during the summers of 1948, 1949, 1950, 1953, and 1954 respectively, for their valuable contributions. Particular thanks are due Dr. Charles F. Deiss, State Geologist, for the opportunity to work on the physiography of the upper Wabash Valley and for permission to publish part of the results of the field work in a non-Survey publication.

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TAIGA-TUNDRA AND THE FULL-GLACIAL PERIOD IN CHESTER COUNTY, PENNSYLVANIA

PAUL S. MARTIN

ABSTRACT. Two shallow cores from a marsh in the unglaciated piedmont of southern Pennsylvania reveal a late Pleistocene pollen sequence. Below 70 cm gritty clays contain abundant sedge, grass and other herb pollens with a tree-shrub pollen sum of 25 to 50 percent. In order of their abundance the arboreal pollens include jack pine, spruce, fir, birch and willow. This flora represents taiga-tundra vegetation of the last glacial maximum, the first pollen evidence of a full-glacial period in North America. Its modern analogue is the subarctic lysotundra, a region of scattered trees in valleys surrounded by bare solifluction slopes.

Radiocarbon samples were collected at 100 cm. in gritty clay and at 150 cm. in coarse sand and clay. The C-14 age of each, approximately 13,500 years, is somewhat younger than the accepted age of the last Wisconsin maximum. Surface contamination of sandy layers, indicated by the pollen diagram, may account for the discrepancy.

The full-glacial clays and sands are overlain by shallow, more organic late- and postglacial sediments. A pine zone and a deciduous forest zone resemble the postglacial sequence north of the glacial border. However, a spruce-fir zone is absent and late-glacial forest invasion in the Pennsylvania piedmont appears to have been of pine rather than spruce.

The biogeographic theory which holds that temperate forest occupied favorable habitats close to the Wisconsin ice sheet finds little support in the Pennsylvania pollen diagrams.

INTRODUCTION

That tundra accompanied glaciation in North America has been a geological postulate, largely lacking biological proof. To the inevitable question, "Did tundra border the ice sheets of Pleistocene times?" Potzger as recently as 1951 found the pollen record "disturbingly silent" (p. 131). Thus we are indebted to Deevey (1951) both for breaking the silence and for indicating how European techniques could be adapted to the collection and preparation of late-glacial sediments.

In northern Maine Deevey encountered high percentages of non-arboreal pollen from lake clays underlying post-glacial marl and gyttja. Not one but two separate "treeless" intervals appeared, representing tundra or taiga conditions. The first accompanied deglaciation while the second marked a subsequent readvance of ice, presumably the Valders (Deevey and Flint, 1957). The studies of Anderson (1954), Leopold (1955, 1956b), and Davis (1957) in Michigan, Connecticut, and Massachusetts respectively, amplify our understanding of the late-glacial environment. Although chronological details remain to be solved, the biological evidence of a late-glacial tundra[1] and taiga now rests on a firm basis.

Older pollen zones in Europe, the full- or pleniglacial, record a time of very high herb pollen percentages (van der Hammen, 1951). Extensive frost-debris and loess-tundra bordered the ice margin during the height of the last glaciation (Frenzel and Troll, 1952). Periglacial features are abundant throughout France (Tricart, 1956) and extend at increasing altitudinal levels from southern Germany and Italy to Ethiopia (Büdel, 1953, 1954).

[1] M. B. Davis and W. H. Drury, who were kind enough to comment on the present article, challenge this assertion. Dr. Davis would define tundra in a floristic rather than a structural sense and reserve judgement on the climatic meaning of late-glacial pollen zones.

In unglaciated eastern North America the geomorphic evidence of cryoplanation and periglacial solifluction includes the following: involutions, ventifacts and frost wedges in New Jersey (Wolfe, 1953; Hack, 1953); boulder fields on steep slopes in Virginia (Nickelsen, 1956), on gentle slopes in Central Pennsylvania (Smith, 1953) and southern Wisconsin (Smith, 1949); a variety of periglacial features in Pennsylvania, including deep surface mantles, boulder stripes, boulder rings and loess (Peltier, 1949; Denny, 1951 and 1956).

It would be satisfying to claim that a deductive search based on these literature clues led to the discovery of an apparent full-glacial pollen record. In reality the following report represents the by-product of a more modest endeavor, to determine the history of a small upland marsh in the unglaciated Pennsylvania piedmont.

The high frequency of herb pollens with small amounts of spruce and pine (figs. 6-9) resembles the late-glacial taiga-tundra spectra of Deevey and others. However, the presence of such a spectrum in an unglaciated area requires us to postulate an age slightly greater than that of the late-glacial pollen zones, the latter deposited during deglaciation. Apparently this is the first North American pollen evidence of a "treeless period" analogous to the European full-glacial.

The interpretation of the Marsh profiles is, however, not simple. In general features they bear little immediate resemblance to published American pollen diagrams. The postglacial portion is greatly compressed and perhaps truncated. There is a superficial but confusing resemblance between the herb zones of the full-glacial (F zones) and the postglacial (C zones). In addition the anomalous mixture of a tundra-taiga pollen spectrum with small amounts of oak, hemlock, hickory and other temperate forest trees complicates the analysis of the F zones. This should not obscure the main point, that a "treeless" interval, climatically controlled, produced the pollen spectrum recorded in the deeper clays.

Beyond a description of the following pollen diagrams I wish to consider their implications. What age do they represent? How extensive are periglacial solifluction deposits and where can similar alluvial-colluvial sediments be sought? How are we to interpret the high percentages of NAP (non-arboreal pollen)? Do these percentages derive from a completely treeless tundra, or can we atribute the small amounts of conifer and birch pollen to a scattering of trees in a taiga formation?

Finally we may consider the matter of vegetation and faunal displacement during the glacial maximum, a classic problem in North American biogeography.

ACKNOWLEDGMENTS

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My best thanks go to Johs. Iversen of the Danmarks Geologiske Undersøgelse. He first recognized the significance of the Marsh sediments and offered patient help with many problems.

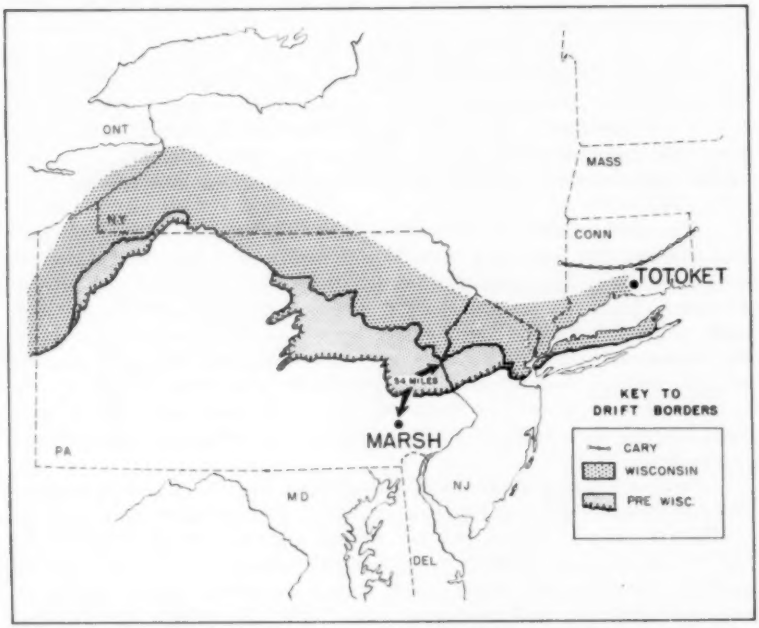

Fig. 1. Location of the Marsh with regard to the Pleistocene drift borders. Totoket, a late-glacial locality (Leopold, 1955, 1956b), lies within the Wisconsin drift border.

LOCATION

Pollen cores M1 and M2 were collected one mile south of the village of Marsh, East Nantmeal Township, Chester County, Pennsylvania (figs. 1 and 2). The locality lies within 54 miles of Wisconsin drift and much nearer to older (Illinoian?) drift (fig. 1). The swales of sedges, ferns and composites which extend for two miles along the Marsh Creek (and give the stream its name) reach a maximum width of one-half mile (figs. 2 and 3). The dark portion of the aerial photograph marks the course of the marshlands with arms

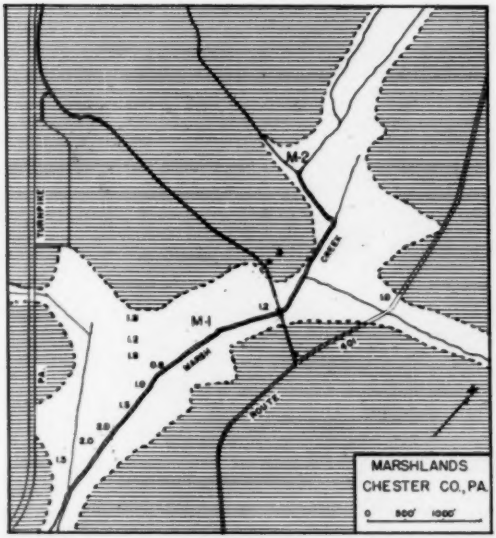

Fig. 2. The headwaters of the Marsh Creek showing location of cores M1 and M2. Uplands are shaded; poorly drained marshlands of the Marsh Creek valley are shown in white, with sediment depth, as determined by soil auger probes, measured in meters.

extending along tiny tributaries. A small artificial dam ponds the southern end. Below this point the Marsh Creek flows at a steeper grade, as shown on the stream profile of figure 4. At Dorlan, about five miles downstream, it joins the Brandywine.

If the contours shown on the U.S.G.S. Phoenixville Quadrangle are approximately correct, the fall in the Marsh proper is 25 feet in two and one-half miles. A more typical profile for a piedmont stream is that of the adjacent South Fork of the French Creek (fig. 4).

Bogs, undrained basins, and *extensive* marshes or swamps do not exist in the unglaciated Piedmont. On the other hand, many small sedge-grass swales mark headwater valleys along the drainage divide separating the Brandywine and Schuylkill watersheds. The Marsh is one of the largest of these, only a few of which appear on the topographic maps of southeastern Pennsylvania.

GEOLOGY AND SEDIMENTS

The portion of northern Chester County including the upper part of the Marsh Creek lies in the Welsh Mountain Anticline. This system of Precambrian sedimentary and igneous rocks in the Piedmont Upland is bordered to the north by extensive Triassic beds of the Newark group. To the south lie the Ordovician and Cambrian limestones of the Chester Valley. The Marsh itself is underlain by Pickering gneiss and quartz monzonite (see Bascom and Stose, 1938).

Fig. 3. Aerial photograph of the headwaters of the Marsh Creek with the portion mapped in Figure 2 enclosed in white dots. North is at the top.

Within the Precambrian rocks are small scattered lenses of Franklin limestone, "associated in every case with the graphite-bearing beds of the Pickering gneiss". Bascom and Stose (1938, p. 23) locate 11 such lenses which form a white, coarsely crystalline limestone or marble. Unlike the limestones of the Chester Valley, which evidence karrenfeld (Bascom and Stose, 1938, plate 10) the Franklin limestone is not a source of sinks or basins. In view of claims made elsewhere regarding poorly drained surfaces in the unglaciated Appalachians it seems important to establish that the Marsh and other flat valleys in northern Chester County are not solution features.

As a part of the Quaternary system Bascom and Stose discuss terrace gravels of the Schuylkill River (1938) and the lower Brandywine (1932). Although they ignore the Marsh, a similar swale, Pine Swamp, is identified

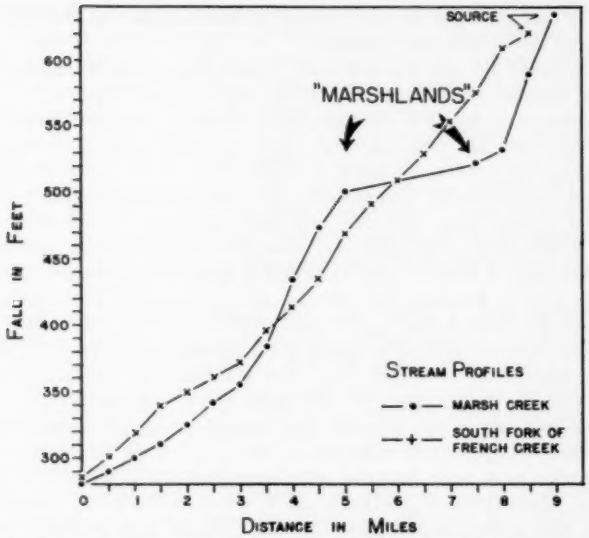

Fig. 4. Stream profiles of the Marsh Creek and its neighbor across a drainage divide, the South Fork of French Creek. The latter joins the North Fork of French Creek at about the same elevation as the Marsh Creek joins the Brandywine.

(1938, p. 85) as alluvium which obscures structure adjacent to Triassic rocks in the Hopewell area. They do not comment on its origin. The late Pleistocene age of alluvium along small Chester County streams is established by the discovery of mammoth and mastodon teeth from the Hockesson Valley and White Clay Creek respectively. The mammoth tooth "was found some five or six feet deep in a meadow, under the black mud, as we call it, in a gravel stratum" (Michener *in* Futhey and Cope, 1881, p. 441). The mastodon tooth was excavated with debris exposed near Avondale following a flood of the White Clay Creek.

A photograph (Bascom and Stose, 1938, plate 9) of a "field strewn with residual Triassic diabase boulders" is typical of many Chester County pastures. Although such residual blocks might be attributed to differential erosion unrelated to Pleistocene climatic change, the dense aggregations of equally massive blocks forming true boulder fields are almost surely of Pleistocene age. These occur commonly through upper Chester, Montgomery, Bucks, Berks and Lehigh Counties and are rare or absent in southern Chester County. Locally they are distinguished as "devil's rocks", "blue rocks" or "ringing rocks". The nearest to the Marsh lies about three miles farther north, one-half mile east of Elverson.

Before pollen analysis revealed the age of the shallow gritty clays, a vain effort was made to locate a "deep" and thereby "old" deposit. Depths reached by soil auger probes are shown in figure 2. An impenetrable bottom, presumably bedrock, invariably lay below 2.5 meters. The auger probes disclosed

gritty clay and sand but no peat. Except in late summer the Marsh is usually wet and the clay is soft and yielding.

From cores M1 and M2 and from a trench dug near M1 during the collection of samples for radiocarbon dating, four sediment types were identified (see signatures on figs. 6 to 9). Core M1 includes: (1) a basal greenish clay, yellow when oven dry, with scattered coarse grit and sand increasing towards the bottom; (2) a wet sandy layer, possibly a lens, of dirty yellow color; (3) a gray gritty clay similar in texture to (1); (4) a brown fine peaty clay with occasional quartz chips, charcoal, and plant remains; (5) extending to the surface, a fine, black, humified clay charged with minute charcoal particles and almost free of sand or grit. Core M2 is similar but does not include the basal clay nor the humified surface clay and charcoal. M1 was collected from the surface, M2 from below a very small pond roughly 50 cm in depth.

Carbon content of samples taken at 25 cm intervals from M1 is shown in figure 5. Apparently the upper gritty clay at 82 and 107 cm is slightly more organic than the deeper clays, but the main rise in organic carbon occurs at the 58 cm level which separates the pine from the oak-pine pollen zones. Pollen and charcoal appear the main sources of carbon in the deeper clays; seeds, needles and coarse plant material with large charcoal fragments feature the pine zone, while charcoal permeates the C zone (above level 57 cm) to such a degree that pollen identification proved difficult. Fire probably occurred throughout the history of the Marsh. The greater accumulation of charcoal in the upper sediments is certainly related to the decrease in inorganic deposition during the postglacial period; it may or may not reflect an increase in fire frequency.

In brief the Marsh valley presents two major geomorphic problems: (1) to explain its unusually flat headwater profile, in contrast to other piedmont streams, and (2) to date its sedimentation. In the absence of limestone bedrock, solution and collapse must be ruled out. According to Bascom and Stose (1938) the Pickering gneiss, which underlies most of the Marsh, contacts quartz monzonite near the lower end of the valley; differential erosion of the softer gneiss might account for the flat profile.

Regarding sedimentary history the pollen evidence points to deposition during a cold, relatively treeless interval. The suggestion that the piedmont block fields were formed contemporaneously is perhaps gratuitous. While I infer that the inorganic clays and sands of the Marsh were deposited largely as a result of periglacial solifluction, details of the process are unknown. Sedimentation study of the Marsh Creek alluvium has been undertaken by J. H. Moss and his students.

VEGETATION AND BIOTA

Braun (1950, p. 247) includes Chester County and southeastern Pennsylvania in the Piedmont Section of the Oak-Chestnut Region. Isolated farm woodlots and small tracts of relatively undisturbed forest on private estates represent the remnants of Penn's Woods. On the most favorable upland sites with northerly exposure the dominant trees include beech, white oak, red oak and tulip poplar. Here the vernal geophyte flora is rich and grows in deep

humus. Hemlock and sugar maple, two common associates of beech in southern New England, are seldom part of the upland mesic climax in the piedmont. Hemlock is confined to steep wet hills along the Brandywine while sugar maple "is not common in Chester County; but it occurs frequently near the forks of the Brandywine" (Darlington, 1837, p. 246).

Upland forests with beech comprise a relatively small part of the regional vegetation mosaic. Braun's effective description of Westtown woods shows that even very gentle changes in exposure can decrease the amount of beech and favor various oaks. On steep south slopes the dominant species include black, post, chestnut and blackjack oaks with an understory of sassafras, chestnut sprouts, sour gum, dogwood and ericaceous shrubs. The mull humus geophytes are scarce or absent.

McClune, as quoted in Futhey and Cope (1881, p. 340) leaves us with the interesting thought that pre-Colombian disturbance of the forests by Indians exceeded that of settlers during Colonial times.

> When the country was first settled the forests were far from dense, the Indians having kept the timber from growing by fire on the high ground for greater facility in pursuing the deer, and in the low grounds and valleys to enable them to hunt the buffalo . . . After the Indians had retired the growth of timber was rapid, and about the period of the Revolutionary war the forests were dense.

Possibly Indian burning accounts not only for the high frequency of charcoal fragments in the Marsh sediments but also for the relatively great abundance of pine pollen in the C zone. Despite the well-known over-representation of pine pollen in most profiles, the genus appears far too scarce near the Marsh today to account for 40 percent of the forest pollen rain. Fire is a primary agent in promoting pine growth. If McClune's estimate of prehistoric burning is approximately correct, this may explain the postglacial pine curve in M1-C.

In addition to the forest communities, serpentine barrens comprise a small but important structural feature of Chester County. These natural savannas extend in a broken belt from Oxford to Malvern. In prehistoric times they were an important habitat for certain prairie animals and plants, probably including the heath hen (*Tympanuchus cupido*) according to Dr. R. B. Gordon (pers. corr.). While most of the barrens are treeless and support *Andropogon scoparius*, *Phlox*, and other herbs, there are scattered groves of low oaks, pitch and virginia pine.

Holding to the climatic climax concept, Braun (1950) suggests that although succession is immeasurably slow, the black and white oak woodland of adjacent slopes will eventually establish itself on the barrens. In view of the geological evidence which indicates that the serpentine has been exposed and available for plant occupation at least since the evolution of the angiosperms, such succession seems possible only in theory.

Although the barrens are of considerable floristic interest, there is little reason to believe that they have contributed in an important way to the Marsh pollen rain. During prehistoric times these areas were separated by a minimum of ten miles of continuous forest.

In addition to the serpentine barrens the Marsh itself, and probably other smaller wet swales, comprised natural openings in the primeval forest mantle.

The high *over-representation* of NAP and fern spores through the C zones provide paleoecological proof of an enduring sedge-grass-fern-composite cover. On the other hand forests may have covered the site, or part of it, during the pine period (B zone), when sedge and grass pollen is minimal, compositae and ferns are virtually absent, and large chunks of charcoal and occasional needles appear in the peaty clay.

The present surface supports patches of tussock sedges, sensitive ferns, composites (especially *Bidens*), scattered cattails and jewel weed (*Impatiens*). Common shrubs include swamp rose, button bush, red osier, alder and poison sumac. The marginal trees are mainly pin oak (*Quercus palustris*) and elm. While I have not attempted quantitative comparison, it is obvious that Pine Swamp is more thoroughly invaded by shrubs than the Marsh.

The naturalist in search of the unusual is attracted to "atypical" habitats such as the Marsh. Regarding the modern flora, Stone (1945) comments that "The swamps along Marsh Creek were, and still are, the home of many plants rare elsewhere in the county . . . " (p. iv). *Evonymus atropurpureus, Hypericum virginicum, Calamagrostis canadensis, Eriophorum tenellum* and *Carex comosa* are among the "rare" species listed by Stone. The disappearance of *Phragmites communis* (reed grass) and *Vaccinium macrocarpa* (bog cranberry), both familiar to Darlington, is not readily explained.

Although the Marsh includes large populations of swale plants which occupy less space in the other County marshes, it does not appear to harbor an endemic flora. Smaller swale habitats which Stone mentions, such as Oldmixon and Brandamoor, have equally rich or perhaps richer floras. I found no species in Stone's Flora exclusively confined to the Marsh.

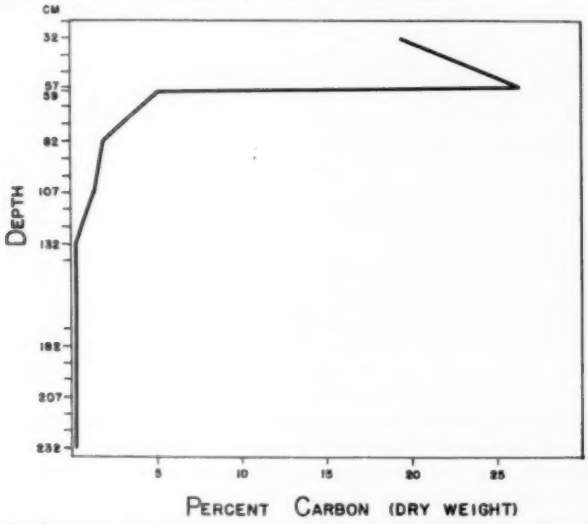

Fig. 5. Percentage carbon content of core M1, determined by chromic acid extraction at the indicated levels.

To students of avifauna the Marsh is an important habitat for various rails, the American bittern, the alder flycatcher and, in migration, the Wilson's snipe. These are rare elsewhere in Chester County (Conway, 1941).

PROCEDURE

Field Work.—Sediment cores M1 and M2 were collected with a Hiller borer and stored under refrigeration in their aluminum liners. Avoiding material at the ends of the liners I removed three samples from each, at intervals of 8 to 9 centimeters. Core M1 is incomplete; the gap between levels 145 and 174 represents an aqueous gritty sand which slumped and was lost. Core M2 penetrated gritty sand, but could not be driven into the deeper clay layer, if one occurs in that part of the Marsh. While manual drilling in heavy, gritty clay is inefficient, I feel reasonably certain that M1 includes all the sediment lying above bedrock.

Laboratory.—The preparation of pollen samples followed the method of Faegri and Iversen (1950), i.e. deflocculation in KOH, 15 to 30 minutes of gentle boiling in hydrofluoric acid, followed by acetolysis. Centrifuge tubes machined from the plastic "Teflon" obviate the use of crucibles, save time, and diminish the danger of losing part of the sample. Hydrofluoric acid added to the sample may be boiled in the "Teflon" tube in a paraffin bath at 170°C. Up to one cc of sediment was required to concentrate sufficient pollen for a rich preparation. All samples were mounted in glycerine and stained with safranine.

Analysis.—Counts of 600 or more pollen grains (excluding spores) at 24 levels were made from core M1 and counts of 300 or more grains at 14 levels from M2. For routine identifications 250x magnification of a Zeiss binocular microscope proved effective; oil immersion aided in more difficult cases. In addition to the illustrations and keys of such standard works as Erdtman, Woodhouse, and Faegri and Iversen, I resorted to a small collection of reference slides. However, a variety of pollen types were relegated to the "Unknown" category. None of these were abundant and I assume that most of them represent various non-arboreal grains, e.g. Rosaceae, Ranunculaceae, Umbelliferae, etc., which contribute to the richness of late- and full-glacial floras. All unknowns were included in the NAP curve.

Generic determination of most tree and shrub pollens proved simple. The recognition of *Populus* is not, however, and I made no consistent effort to separate this from the rather similar sedge grains. Svend Anderson, who is very well acquainted with the American poplar-sedge problem, found two poplar grains and 34 sedges from level 140 of core M1. Other tree species present in the Marsh cores which were not systematically counted included *Larix* and *Juniperus*.

THE POLLEN DIAGRAMS. I. ZONAL ANALYSIS

The late- and postglacial pollen sequence developed by Deevey, Leopold, and others for northeastern United States is summarized by Deevey and Flint (1957) in table 1.

Using this chronology, which admittedly applies only to the glaciated section of eastern United States, it is of interest to determine how the Marsh

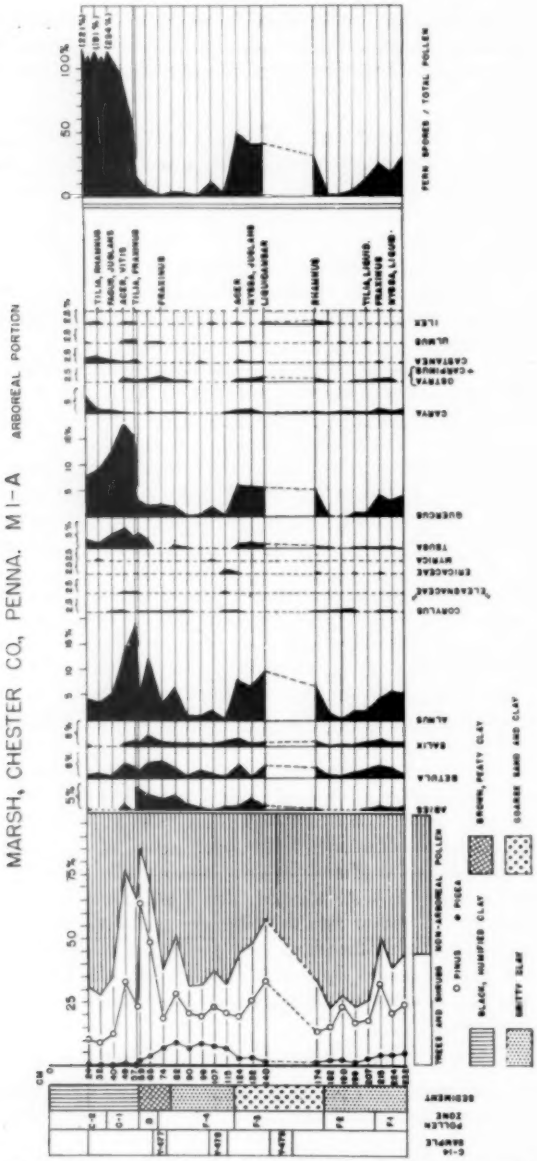

Fig. 6. Pollen diagram M1-A. Approximate depth of C^{14} samples, pollen zones, sediment types and sampling levels are shown along the left margin. The non-arboreal pollen sum (NAP) does not include fern spores.

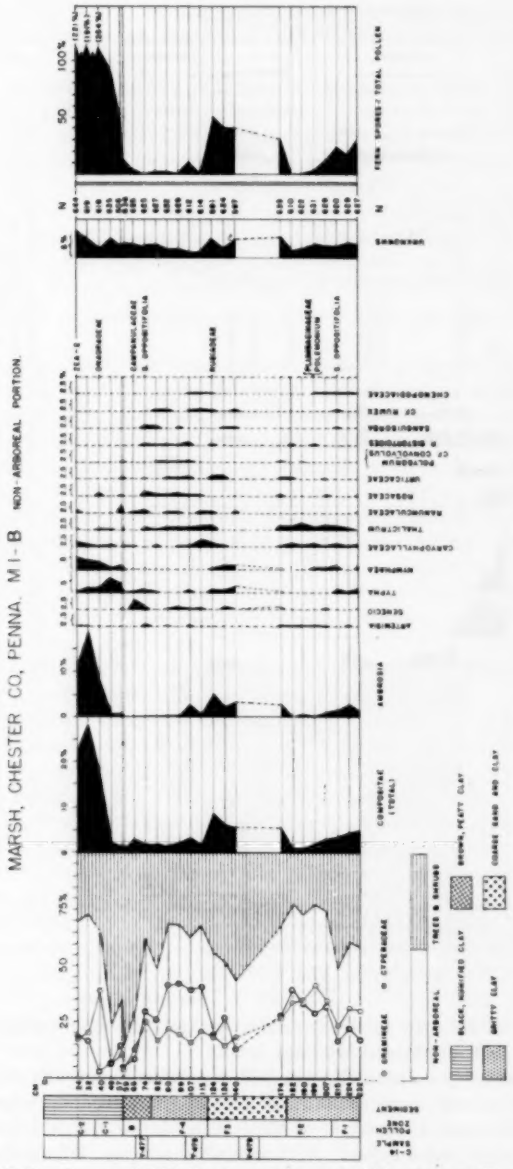

Fig. 7. Pollen diagram M1-B. See legend figure 6. Abbreviations: P. = *Polygonum*; S. = *Saxifraga*. For *P. convolvolus* read *P. convolvulus*.

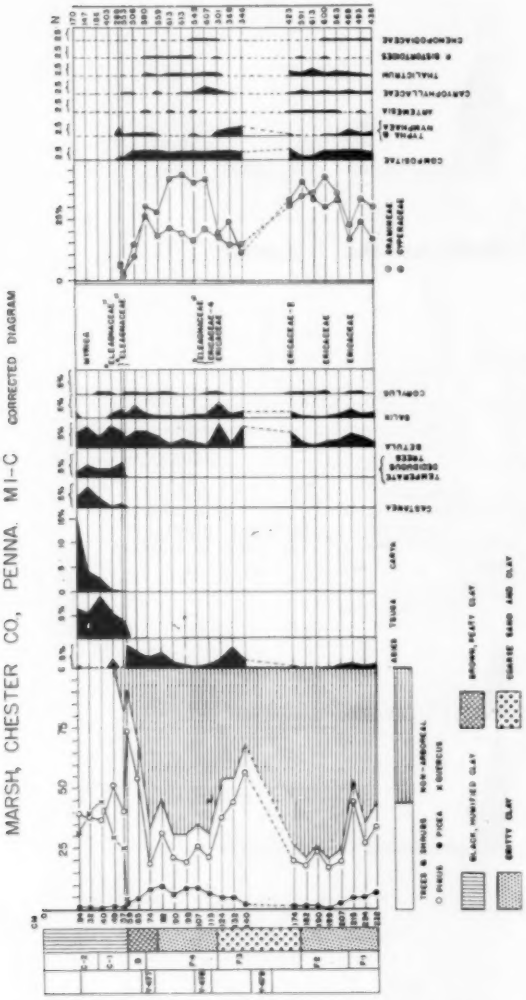

Fig. 8. Pollen diagram M1-C. See legend figure 6. Subtractions applied to this diagram are as follows: all (NAP) above level 57; oak, alder and all temperate forest pollens below level 59; Compositae and other NAP below level 59 reduced on a percentage basis (see text). The corrected diagram is an attempt to obtain the regional pollen rain, undistorted by NAP over-representation in the upper portion and by surface contamination of deeper sandy levels in the lower portion of the profile. Abbreviation: P. = *Polygonum*.

diagrams will "fit". An obvious starting point is the pine period (B zone), well marked in both M1 and M2 (figs. 6-9).

The B zone (pine period).—In both diagrams pine is the dominant pollen; spruce is scarce; hemlock and oak are important secondary species; NAP reaches a minimum, 15 percent. In these features the Marsh diagrams match closely Deevey's and Leopold's Connecticut diagrams. While chronological equivalence is doubtful, they offer a convenient stratigraphic landmark. The

Marsh Pine Zone should be older than its stratigraphic equivalent in New England and may represent late-glacial rather than postglacial events (Valders time and older).

Pollen percentages obtained from a single sample of the Pine Swamp alluvium, three miles north of the Marsh, are as follows: *Pinus* 38, *Picea* 3.4, *Abies* 2.8, *Tsuga* 1.9, *Betula* 5.6, *Quercus* 2.2, *Alnus* 10.2, *Salix* 0.9, *Carpinus-Ostrya* 1.2, *Castanea* 0.3, Ericaceae 0.1, Gramineae 6.2, Cyperaceae 14, *Nymphaea* (?) 7.4, Compositae 0.9, *Typha* 0.3, *Sanguisorba* 0.3, unknowns 3.4, Σ AP 67.2 percent. The sample represents compact black clay from the bottom of a 125 cm core. I had expected an equivalent of the Marsh F zones in the pollen spectrum; however, the percentages fit closely the B zone, especially level 87 of M2. Unless the collecting data are erroneous, the Pine Swamp alluvium contains a much longer postglacial sedimentary column than the Marsh.

The C zones (oak, hemlock and oak, hickory).—Despite the high percentage of herbs, the C zones in M1 and M2 are clearly a postglacial record. There is no indication in the pollen profile that the climate was colder than at present. In an effort at recovering the upland (forest) pollen rain I have subtracted extraneous herb and alder percentages in levels 24 through 49 in diagram M1-C. Oak and pine predominate, together averaging 75 to 80 percent. Hemlock, hickory and chestnut are of lesser importance. The increase in hickory and slight decline of hemlock suggests that zone C-2 is drier than zone C-1 and thus represents the thermal maximum or xerothermic, frequently encountered in postglacial diagrams. Zone C-3, the oak-chestnut zone of Deevey and Flint, should be sought in the surface layer of the Marsh, which was neglected during field work.

The Pennsylvania C zones differ from the Connecticut sequence, and resemble diagrams from Northern New Jersey (Potzger and Otto, 1943) in their higher percentages of pine and chestnut and virtual absence of beech.

While the pollen percentages for trees in profiles M1 and M2 are rather close, there is considerable difference in the values of NAP and alder. Alder is more abundant in M2 while composites, especially *Ambrosia*, and fern spores are more so in M1. High fern, alder and composite percentages accompanied by abundant grass and sedge pollens can be regarded as the result of *local over-representation*, one of three "classic sources of error, which have been known and feared during the whole history of pollen analysis" (Faegri and Iversen, 1950, p. 91). In slowly accumulating subaerial marsh or spring seep sediments, as opposed to lake gyttja or peat, the regional pollen rain is partly distorted by locally growing NAP and fern spores. In C-1 and C-2 of Leopold's Durham Meadows core A the NAP averages about fifty percent. Various Danish postglacial diagrams illustrate this effect (see Iversen, 1954).

The A zone (spruce, fir).—An extraordinary feature of the Marsh pollen profiles is the absence of a forest pollen zone dominated by spruce. A spruce peak, 19 percent at level 95 in M2, may mark the beginning of a spruce period, but the high value of NAP at this level, 60 percent, certainly does not suggest the advance of forest. Unless the high NAP percentages of the F zones represent local over-representation, the late-glacial forest invasion in south-

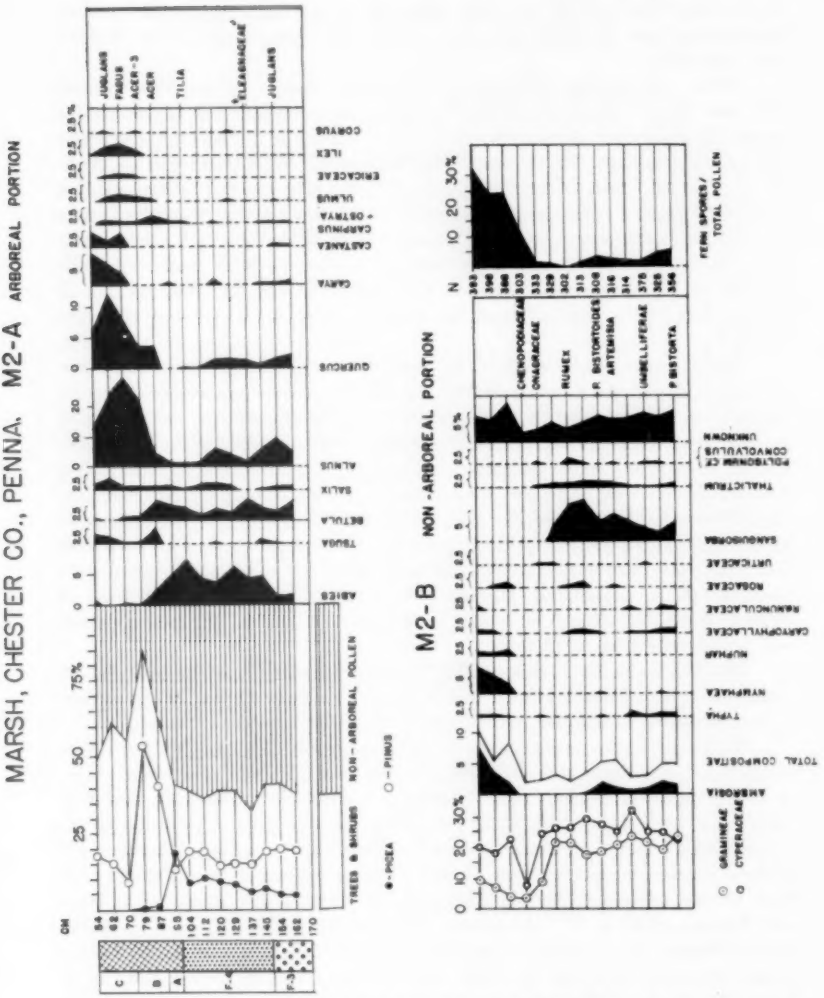

Fig. 9. Pollen diagram M2-A. See legend figure 6. Abbreviation: P. = *Polygonum*. For *P. bistorta* read *P. bistortoides*.

eastern Pennsylvania was of pine, not spruce, and occurred quite rapidly. From a climate *too cold* for forest (the F period taiga-tundra) there was a rapid change to one *too warm* for extensive spruce forest (the B pine period). The pine invasion was marked by a sharp rise in large-sized pine pollen, i.e. species other than jack pine (figs. 10 and 12).

TABLE 1
Late- and Postglacial Pollen Chronology
(from Deevey and Flint 1957)

Zones	Pollen Spectrum	Age in years B.P.	
C-3	Sub-Atlantic; oak, chestnut	0 - 2000	POST-GLACIAL
C-2	Sub-Boreal; oak, hickory	2000 - 5500	
C-1	Atlantic; oak, hemlock	5500 - 8500	
B	Boreal; pine	8500 - 9500	
A	Pre-Boreal; spruce, fir, pine, oak	9500 - 12500	LATE-GLACIAL
T-3	Younger Herb Zone; park tundra	12500 - 13500	
T-2	Pre-Durham Spruce; spruce, pine, birch	13500 - 13800	
T-1	Older Herb Zone; tundra	13800 - 15000	

The F zones (full-glacial herb zones).—Below the pine period in M1 are four distinct NAP or herb zones. Ignoring for the present the problems raised by F-1 and F-3, we may consider first the spectrum of F-2 and F-4. In both the NAP percentages reach a maximum point; pine, spruce and fir are the only important trees; there is no indication of contamination. The percentages of fern spores, composites, alder, oak and temperate forest pollens are minimal. We may take the pollen rain as representing the true upland vegetation of the piedmont. F-2 has decidedly less spruce and fir with a greater amount of herbs (77 percent in level 182) than does F-4. It may be slightly colder. While not a true tundra, these F zones represent a taiga-tundra with widely scattered spruce and pine in a grass-sedge steppe.

Pollen preservation, as measured by the ratio of broken to entire grains (fig. 12), is better in F-2 and F-4 than in the rest of the profile. Size frequency of pine wings indicates a small species, almost surely jack pine, *Pinus banksiana*, (fig. 10). *Thalictrum, Artemisia* and *Sanguisorba canadensis* reach a maximum frequency while *Polemonium* and *Plumbaginaceae* were not found elsewhere.

It is inconceivable that such an assemblage grew here in postglacial time. The late-Wisconsin glacial maximum is the most probable period of deposition for the F zone clays. On the other hand this does not preclude formation of at least part of the deeper clays during an earlier glaciation, or during the early Wisconsin. Taken at face value the F-2 and F-4 pollen profiles indicate a rela-

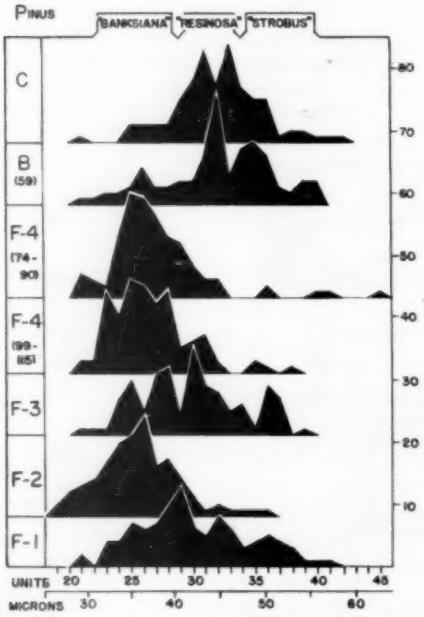

Fig. 10. Size-frequency measurements of 100 *Pinus* wings from each pollen zone in core M1. Jack pine (*P. banksiana*) is provisionally assigned to the smallest size class; larger modes may represent *P. resinosa* and *P. strobus*.

tively treeless vegetation of a cold interval. They do not indicate which glacial advance produced the environmental change.

The possibility that *local over-representation* is responsible for the high NAP values in these F zones may be rejected on two counts: first, there is no rise in composite and fern curves, both typical features of postglacial over-representation; second, the highly inorganic character of the sediments with well preserved, unbroken pollen grains suggests rather rapid pollen capture and deposition. A slowly accumulating sediment with NAP over-representation would be more organic and pollen preservation would be poor.

If the interpretation of Zones F-2 and F-4 is reasonably straightforward, that of F-1 and F-3 is perhaps less obvious. The latter are distinguished by peaks in the curves of alder, oak, hemlock and other temperate forest trees, composites, *Nymphaea*, *Typha* and fern spores. All of these reach values exceeded only in the postglacial sediments. In fact the frequency changes which distinguish F-1 and F-3 bear a remarkable resemblance to the C zones. The relative abundance of the species which characterize Zones F-1 and F-3 is *the same* as their relative abundance in the postglacial (C zones) part of the diagram, i.e. fern spores > Compositae ≅ *Alnus* ≅ *Quercus* > *Typha* ≅ *Nymphaea* ≅ *Tsuga* ≅ *Carya* > *Ulmus* ≅ *Castanea*. Only *Carpinus-Ostrya* pollen has a slightly higher frequency in the F zones than might be expected on the

basis of its postglacial abundance. Apparently F-1, F-3 and, to a lesser degree levels 82 and 107 of F-4, have been contaminated by rebedded material. This is indicated by the presence of small amounts of climatically anomalous species as oak, fern spores and by poor preservation with a higher percentage of broken pine grains than is typical of the F-2 and F-4 zones.

The source of contamination could represent erosion and redeposition of an interglacial sediment. However, the high correlation between the contaminated element in full-glacial sediments and the normal postglacial profile directs suspicion to ground water transport from the surface. While the compact gritty clay of zone F-4 is impervious, as is the brown clay of zone B, the gritty sand of F-3 is quite fluid. Possibly it is in direct contact with the surface at some point. It is of interest that in diagram M2 the coarse sand level (F-3) shows less contamination than its equivalent in M1.

To anticipate the objection that the oak, hickory and hemlock pollens of the F zones represent climatic change and the invasion of a few temperate trees, we may recall the following: (1) there is no indication of increased organic sedimentation in zones F-1 and F-3, a likely feature of major climatic improvement; (2) the invasion of a spruce-jack pine savanna by scattered oak, hemlock and other temperate forest trees would constitute a clisere unknown in the modern vegetation pattern.

In an attempt to recover the natural pollen rain of the F zones, I employed a series of subtractions (diagram M1-C). The following curves were removed: fern spores, alder, oak, hickory, hemlock and other temperate forest trees and shrubs. Compositae were reduced to a maximum of 10 grains per level. Rebedded sedges and grasses, which would go undetected in the analysis, were reduced empirically by an extrapolation of the postglacial NAP/fern ratio, very roughly two to one. This is, of course, too low a ratio for levels 24, 32 and 40; however, it is a close approximation for levels 48 and 57.

The corrected AP/NAP curve obtained in M1-C shows a rise in regional tree pollen, especially pine, and thus implies a "thickening" of the taiga and climatic improvement in zones F-1 and F-3. Unfortunately not until size-frequency measurements were finished did it appear obvious that I had failed to correct for rebedding of pine pollen in these zones. A significant rise in mean and variance correlates closely with the contamination levels, the result of a mixture of large with small sized grains. The former represent rebedded material while the smaller grains are presumably jack pine, part of the regional vegetation of the full-glacial. For this reason no climatic significance can be attached to the apparent rise of pine in F-1 and F-3. The various F zones distinguished in the pollen profile may not represent markedly different climatic intervals.

THE POLLEN DIAGRAMS. II. SELECTED ARBOREAL POLLEN CURVES

Within a genus it is sometimes possible to relate morphological with stratigraphic and climatic events. Although specific identification may be impossible, we can describe changes in size frequency within the genus as a whole and construct a curve showing shifts in mean and variance through time. Unless there has been differential shrinkage in preservation or prepara-

tion of samples, or rapid evolution of pollen size in the recent past, such a curve will record shifts in species composition.

The Pine Problem.—In the words of Darlington, "All the pines are rather scarce in Chester County" (1837, p. 548). Nevertheless, they contributed a fair share to the pollen rain in the postglacial C zones as well as the major share of tree pollen in B and F zones. Four species occupy Chester County today. *P. strobus*, the white pine, grows locally in northern Chester County and forms a rather dense stand at Pine Swamp; *P. virginia* is common on the Serpentine of eastern Chester County, a habitat occupied by *P. rigida* in southwestern Chester County; Stone (1945) has recorded the Table Mountain pine, *P. pungens*, from western Chester County.

Size-frequency measurements on 100 pine wings from each pollen zone are shown in figure 10. While cap length (measured as the distance between wing insertions) is the standard pine measurement, I have preferred wing (bladder) length for various reasons. In the C layers unbroken grains suitable for cap measurement were difficult to find. In the F zones there may be differential breakage of pine grains, depending on size. Measurement of entire grains exclusively would produce an artificially high frequency of small individuals.

The frequency distribution curves show an important shift in size between the various F zones and the postglacial record. I attribute the small size of pine wings in F-2 and F-4 to jack pine (*Pinus banksiana*). Modes larger than 40 microns represent other species. In addition to the pines inhabiting Chester County today *P. resinosa* may be represented.

Measurements used in the size-frequency curves of figure 12 were also analyzed by levels, as shown in figure 11. This diagram follows the method of Hubbs and Hubbs (1953) and makes a quantitative comparison through time in the same fashion that they apply it to the analysis of spatial variation. Those populations in which the black bars (twice the standard deviation of the mean) do not overlap are significantly different at the one percent level. If the overlap of the larger on the smaller bar does not exceed 30 percent, the level of significance is roughly five percent. White bars on either side of the mean represent the standard deviation, a measure of variability. Low variability is encountered mainly in zones F-2 and F-4 where it appears that only a single species is represented.

Figure 11 exhibits an interesting correlation with the contamination of F zones. In those levels with relatively high values of oak and other redeposited material, the mean size and standard deviation of pine increases. This includes levels 82 and 107 of F-4, an otherwise "uncontaminated" zone.

Birch size changes.—Birch is relatively rare in the Marsh profiles. There is no indication of an important invasion of tree birch as might be expected early in late-glacial time. However, with late- and postglacial events compressed into a short sedimentary column, such an event might have been missed in the relatively wide sampling interval.

Leopold (1956a) has demonstrated that the smallest birch pollen in New England is that of dwarf birch, *Betula glandulosa*. Without modern material treated with hydrofluoric acid it is impossible to make valid size-frequency

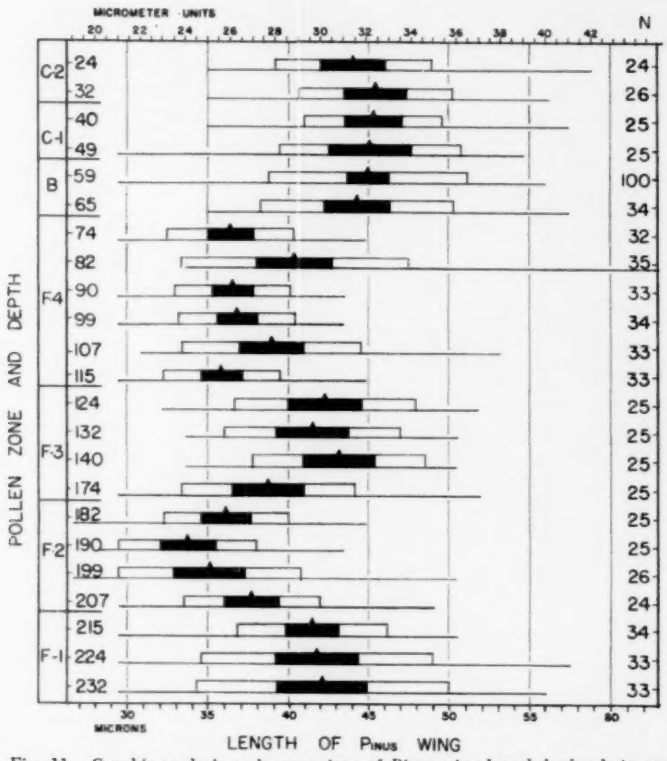

Fig. 11. Graphic analysis and comparison of *Pinus* wing length by levels in core M1 showing mean, standard error of mean, twice the standard deviation and range (see text).

comparisons between the Marsh birches and Leopold's measurements. However, the significant size changes (fig. 13) through various pollen zones are of interest and demonstrate that the F zone birch pollen was smaller than those of the postglacial C zones. Quite possibly dwarf birch is present, at least in F-2. To account for the increased mean in F-1 and F-3 I suspect rebedding, similar to the effect evident in pine size frequency. However, it is not clear why F-4 should have a significantly larger size of birch pollen than F-2. If this represents the invasion of tree birches, it would confirm the evidence from spruce and fir curves that F-4 was slightly warmer than F-2.

Pore Frequency in Alder.—Pore counts on *Alnus* are summarized in figure 16. The three and six-pored grains are relatively rare and are grouped with the more numerous four and five-pored types respectively. Except in the B zone four-pored percentages dominate. Leopold (1955) has compiled pore frequency data on the northeastern alders. Four-pored types include *A. serrulata*, *A. maritima*, *A. rugosa*, and *A. rugosa* var. *americana*. Only the latter

can be considered subarctic; the others grow almost entirely south of the boreal forest.

The five-pored species, *A. crispa*, ranges from southern New England north to Labrador and Alaska. Within the B zone is an increase in five-pored grains which I refer to this species. 36 out of 67 grains at level 79 in M2 and

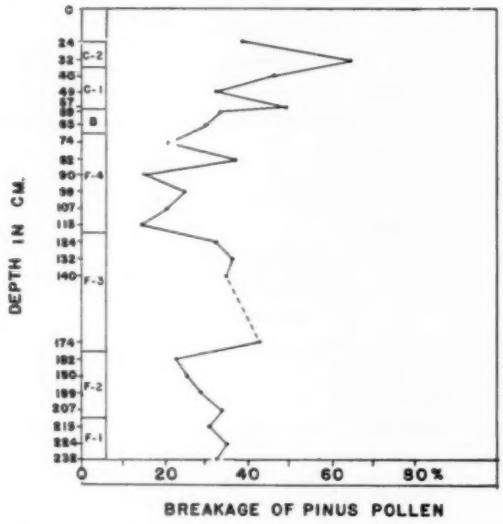

Fig. 12. Percentage of broken *Pinus* pollen computed from the formula $\frac{m}{3n + m}$ where m = number of wings and caps and n = number of entire grains. Preservation is poorest in the surface layers and in the deeper levels contaminated by surface material, i.e. F-3 and F-1.

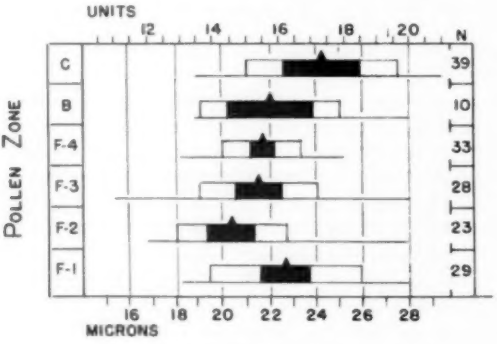

Fig. 13. Graphic analysis and comparison of *Betula* pollen, showing mean diameter, standard error of mean (black bar), twice standard deviation (white bar) and range.

PORE FREQUENCY IN ALNUS

Fig. 14. Pore frenquency in *Alnus* pollen. Percentage of five and six-pored grains shown in black, three and four-pored grains in white.

17 of 37 grains at level 59 in M1 are five-pored grains. The presence of *A. crispa* in the B zone signifies a cool climate, and may allow the correlation of the Marsh B zone with late-glacial events in New England.

Unfortunately, interpretation of the full-glacial alders is impossible on the basis of pore frequency alone. They could represent the boreal and subarctic *A. rugosa* var. *americana*, they could represent rebedded grains from other levels, or a mixture of these two sources. The close "fit" of alder and oak curves through the F zones suggests that the alder is largely rebedded and not a part of the full-glacial vegetation.

THE POLLEN DIAGRAMS. III. IDENTIFICATION OF NAP

In the search for stenotherms of value as climatic indicators one turns to the herb rather than the arboreal pollen types. Although pollen grains lack sufficient morphological detail to allow a complete floristic inventory, it may be possible to determine whether the American periglacial tundra flora was largely derived from temperate sources (Raup, 1951), from arctic-alpine species or from a mixture of both. Intensive study of the Marsh clays should increase the full-glacial floristic catalogue. The following identifications are by no means an index of the total flora.

Thalictrum (fig. 15).—Periporate grains with rather faint pores may represent either *Thalictrum* or *Liquidambar*. In addition to its larger size the latter is distinguished by a pitted surface, clearly evident under oil immersion. *Thalictrum* grains range in size from 14 to 22.5 microns with the mode at 19. In Europe *T. alpinum* is a characteristic member of the full- and late-glacial flora and is the logical species to be expected in the Marsh sediments. A few scattered grains in the B and C zones are of large size, exceeding 21 microns, and probably do not represent the F zone species.

Sanguisorba cf. *canadensis* (fig. 15).—Six-colporate grains with an exine three to four microns thick and measuring about 32 by 21 microns were referred to this species. The fossil grains match closely both *S. canadensis* and *S. officinalis*. Their allocation to the former is based on present distribution and not on morphological separation. *Sanguisorba* is decidedly more common

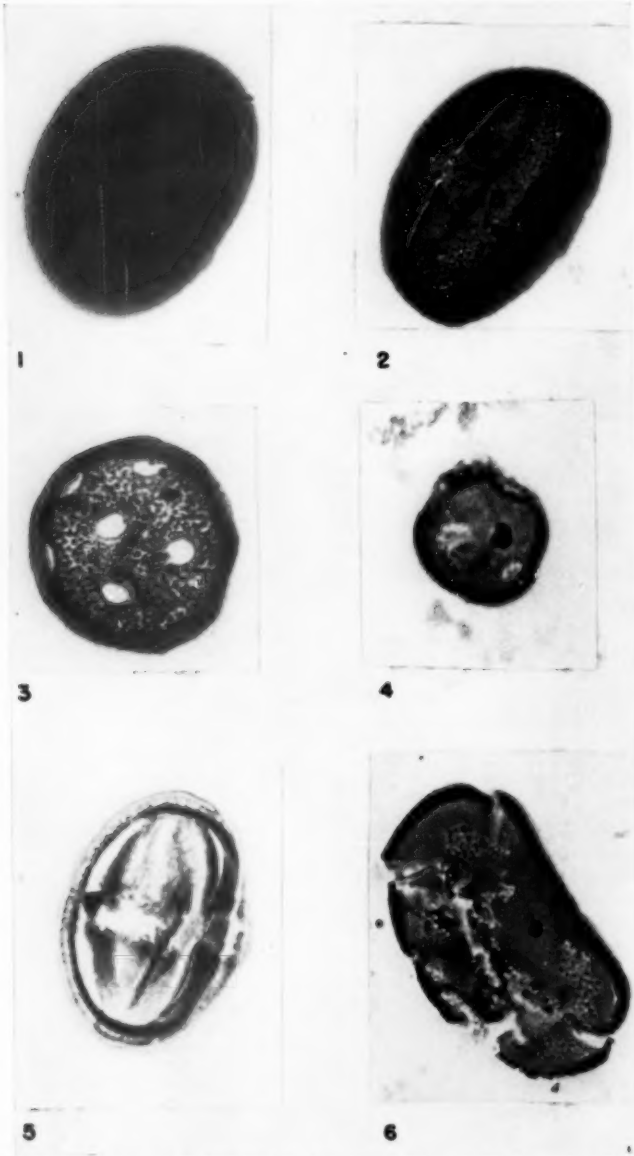

Fig. 15. Photomicrographs of Marsh pollen grains: 1. *Polygonum bistortoides* (modern); 2. *Polygonum* cf. *bistortoides* (fossil); 3. Caryophyllaceae; 4. *Thalictrum*; 5. *Sanguisorba* cf. *canadensis*; 6. corroded fern spore, abundant near the surface and in those deeper levels contaminated by surface material, i.e. F-3 and F-1.

in core M2 than in M1. Despite the presence of *S. canadensis* in the modern Chester County flora, no pollen evidence of it appeared in the C zones.

Polygonum cf. *bistortoides* (fig. 15).—This is a very large tricolporate grain with a small polar area. While the fossil grains, prepared with HF treatment, are somewhat smaller than those on a reference slide of *P. bistortoides* (E. S. Deevey No. 330), they are nevertheless large, measuring 41 to 45 microns in length by 27 to 36 in width with an exine thickness of 4.5 microns at the pole and 3.0 at the equator. *P. bistortoides* occurs "along rills in subalpine meadows" (Fernald 1950) in Newfoundland and in the Cordilleran region. It is the most boreal of the herbs identified in the Marsh alluvium to date.

Polygonum, convolvulus group.—Prolate, tricolporate, psilate grains with conspicuous costae equatoriales were referred to this type. The smallest measured 25 by 10 microns and the largest 30 microns in length. Among American members of the group (*Tiniaria* in Fernald 1950) *P. cilinode* is possibly the species represented.

Plumbaginaceae.—A single intectate, tricolpate grain of very large size, 63 by 51 microns, appeared in level 179 of M1. If Iversen (1940) is correct in assigning all American *Armeria* to the monomorphic type with characteristic ocellate appearance, the present grain must represent another member of the family, possibly *Limonium*.

Polemonium.—This distinctive periporate grain with conspicuous striae was seen only once, in level 179.

Saxifraga cf. *oppositifolia*.—Small tricolpate grains with parallel striae were identified tentatively as belonging to this species. Its distribution is arctic-alpine, reaching northern Vermont.

SUMMARY OF MARSH POLLEN DIAGRAMS

The major features of the Marsh pollen profiles are summarized in figure 16. From the surface downward four distinct pollen zones are present as follows:

1. Temperate forest of oak and pine with smaller amounts of hickory, hemlock, chestnut and other deciduous trees dominates the C zones. Non-arboreal pollen and fern spores are abundant, more so in M1 than in M2. They represent over-representation of locally derived pollen and spores from plants growing on the Marsh surface. Absence of boreal plants indicates a climate subequal to that of the present and despite their brevity the C zones are considered postglacial in age.

2. The B zone is dominated by pine pollen, mainly of large size. The presence of spruce, fir, some small pine grains, i.e. jack pine, and *Alnus crispa* enhances the possibility that this zone is somewhat cooler than the C zones. It may represent late-glacial time in southern Pennsylvania.

3. The majority of the Marsh sediments contain a flora of small pine grains, spruce, fir and abundant NAP. The high NAP percentage in F-4 and F-2 implies a large area of treeless vegetation with elements of taiga woodland along the stream valleys.

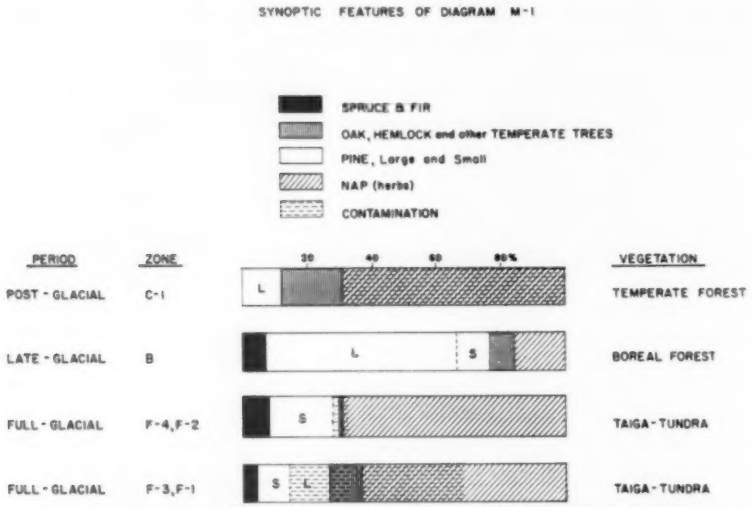

Fig. 16. Synopsis of Marsh pollen zones showing major features of the pollen diagrams. NAP over-representation near the surface and surface contamination in zones F-1 and F-3 is indicated by dashes.

4. Presumably the vegetation of F 3 and F-1 was similar to F-4 and F-2, i.e. a taiga-tundra ecotone. However, rebedded temperate forest and marsh pollen from the surface complicates the analysis. Unless the contamination can be properly recognized and subtracted it is impossible to claim that F-3 and F-1 were climatically distinct from the other F zones.

RADIOCARBON DATING

A preliminary suite of radiocarbon samples (Yale dates, unpublished; see table 2) confirms the pollen evidence that the shallow Marsh clays date from the last glacial (Wisconsin) maximum. While they leave no ambiguity about which glaciation is represented in the upper sediments, these samples are of questionable value in attempting close correlation with Wisconsin stadial events. Y-478 and Y-479 comprise particulate carbon concentrated from a large volume of sediment, roughly six kilograms in each sample. This organic material represents charcoal, pollen and scattered plant fragments trapped in heavy clay and sand. Dating of such material is no more reliable than the assumption of superposition, which is highly suspect in the case of Y-479.

TABLE 2
Marsh C-14 Dates

Yale No.	depth of sample in meters	pollen zone	age, years B.P.
Y-477	0.66-0.76	B	sample insufficient for dating
Y-478	1.03-1.13	F-4	13,540 ± 270
Y-479	1.42-1.57	F-3	13,630 ± 230

While surface contamination almost certainly has made Y-479 "too young", the same criticism carries less force in the case of Y-478. This sample lay in a relatively uncontaminated zone. If its age is approximately correct it raises the extraordinary possibility that jack pine-spruce taiga grew in southern Pennsylvania contemporaneously with the deglaciation of southern New England (see table 1).

TOWARDS A FULL-GLACIAL SYNTHESIS

If not unique, the Marsh pollen diagrams are remarkably unlike published North American profiles. Organic content at all levels is relatively low. The late-glacial and postglacial portion of the sedimentary column is quite short, roughly 50 cm. Problems of absolute dating remain unsolved. Contamination in the form of "over-representation" and "redeposition" complicates the interpretation of certain levels. In begging these difficulties the following discussion may entail more speculation than synthesis. Should it suggest a fresh approach to paleoecological study of the glacial maximum, it will serve its purpose.

Arcto-boreal vegetation and the full-glacial analogue.—Van der Hammen (1951) has examined recent pollen samples from various surfaces in northern Lapland. The NAP totals in three communities follow: (1) middle alpine belt, 83 percent; (2) large open area in the subarctic birch forest belt, 60 percent; (3) subarctic birch forest with scattered pines, 20 percent. The NAP totals of the Marsh F zones, 49 to 78 percent, fall largely between the alpine and subarctic forest percentages. On this basis I interpret the Pennsylvania F zone as transitional between an open boreal woodland or *taiga* and a treeless *tundra*. However, proper evaluation of NAP percentages will require much more study of the modern arctic and alpine pollen rain.

The interpretation of late- and full-glacial profiles may require a brief review of arcto-boreal vegetation. Excluding high arctic desert tundra the three main vegetation zones from south to north comprise the boreal forest, taiga and tundra. These zones fall into three separate biochores, forest, savanna and grassland (Dansereau, 1957) and fit roughly Merriam's life zones Canadian, Hudsonian and Arctic-Alpine.

Of the three taiga or boreal woodland is perhaps the least familiar to American naturalists. In eastern Canada it occupies a belt roughly 350 miles wide between 51 and 56 degrees north latitude (Hustich, 1949, 1951; Rousseau, 1952; Hare, 1950). A subalpine taigoid zone appears on the Shickshock Mountains between 2800 and 3300 feet. In Alaska taiga proper is seldom distinguished (i.e. Drury, 1956) and it is not clear whether there is an equivalent to the Labrador-Ungava taiga.

Dansereau (1957, p. 334) defines taiga as "a needle-leaved (usually evergreen) parkland or savanna", a formulation I have adopted. On the ground that Russian usage is not so specific, including the boreal forest proper, Hare (1950, 1954) prefers the term boreal woodland. Irrespective of terminology Hare, Hustich, Rousseau and Dansereau (1955) are in accord regarding its bioclimatic importance.

As Hare and Taylor (1956) demonstrate, the taiga-forest boundry of southern Labrador-Ungava is relatively sharp. By way of contrast the interdigitation of taiga and tundra zones, controled by the gentle northward dip of the Laurentian Plateau, is so extensive that Rousseau (1952) proposes recognizing it as a separate zone, the hemiarctic. This mixture of taiga and tundra communities is equivalent to the lysotundra of Eurasia (Hustich, 1949; Hare, 1954). The translation "forest-tundra" may be somewhat misleading as no forest, in the strict sense, is present.

Along the Koksoak River valley near Fort Chimo the taiga-tundra ecotone involves scattered small trees of larch and black spruce with thickets of dwarf birch, *Ledum* and other heaths. Very low hills capped with *roches-moutonnées* are treeless and support a shrub-lichen cover. At their altitudinal limit above the valleys the trees are scattered and short, reaching 10 or 20 feet in height. They do not form "Krummholtz". The variety of heaths and heath-like plants (*Ledum groenlandicum, Empetrum, Vaccinium uliginosum, V. vitis-idaea, V. microcarpus, Kalmia, Arctostaphylos alpina*), the relative unimportance of grasses and sedges, the dominance of larch and the absence of jack pine removes any possibility of a close *floristic* relationship between the northern Ungava and the Pennsylvania full-glacial. However, the *structural* resemblance may be close, i.e. scattered trees in the valleys, with bare ridgetops and slopes. If a relative relief of 100 to 300 feet can determine tundra-taiga distribution at the present northern limit of taiga communities, it is probable that similar relief in the Pennsylvania piedmont was sufficient for a full-glacial taiga-tundra ecotone.

The taiga-tundra boundary is of geological as well as ecological interest. In mountainous areas Büdel (1944) locates the lower limit of modern solifluction in the scrub-forest belt, a little above tall timber. Büdel's "scrub-forest belt" may represent the subalpine equivalent of the sub-arctic taiga. At Knob Lake, Quebec (latitude 55 degrees north) the alpine lichen-heath ridges above 2300 feet exhibit a variety of patterned ground features as stone rings, stone stripes, frost boils, asymmetrical valleys, boulder-paved ponds and nivation hollows (Derruau 1956, Henderson 1956). Below 2300 feet within the taiga patterned ground is limited to occasional frost polygons which break the lichen mantle.

Patterned ground and frost action are thought to restrict tree distribution in the Arctic (Raup, 1951). In the Yukon, Denny (1951) found that "most congeliturbation and related frost processes take place in areas devoid of forests." On the other hand Drury (1956) notes that certain solifluction features as string bogs or *strangmoor* appear in the boreal forest. Sigafoos and Hopkins (1951) attribute tussock formation in New England to frost action. With this precaution we may interpret most of the midlatitude "periglacial phenomena" as fossil evidence of tundra and a tundra climate.

Piedmont swales, glades, balds and subalpine parks.—The above brief discussion of modern taiga-tundra and solifluction features constitutes a point of departure for considering the glacial maximum. We should anticipate finding solifluction deposits at appropriate altitudinal-latitudinal levels throughout unglaciated America. In the central Appalachians, treeless swales or "glades"

may be the product of full-glacial climatic change. Cranberry Glades, West Virginia, elevation 3375 feet, is the largest of several similar bogs lying roughly 150 miles south of the Wisconsin drift border (Darlington, 1943). Clays and inorganic sediments reach a depth of 16 feet. They are partly overlain by peat, algal ooze and other organic material. The surface is imperfectly drained by an underfit stream, falling 45 feet in three miles, roughly equivalent to the slope of the Marsh valley. A radiocarbon date of the pine-birch maximum, C-336, 9434 ± 840 years, near the bottom of a pollen profile establishes organic sedimentation as mainly postglacial.

Regarding its erosional history, Darlington was confronted " . . . with the puzzling question of the origin of a wide valley so near the source of a stream. This is no ordinary occurrence" (1943, p. 375). In the absence of surface limestone beds he ruled out origin by sinking. Stream erosion during Appalachian peneplain cycles could account for the gentle profile of the upper Cranberry River, but would not explain the change from degrading to aggrading action. "This could have happened possibly during the latter part of the glacial period, when increased precipitation, decreased evaporation, and lesser vegetation cover on the slopes caused the swollen streams to be overloaded with sediments, which were deposited over the level area" (Darlington, 1943, p. 384). Although Darlington's interpretation foreshadows present periglacial theory, pollen analysis of the deeper clays is needed to resolve the question of whether climatic cooling as well as increased precipitation reduced the vegetation cover.

At Glade Bog, 2400 feet, in southwestern Pennsylvania, Schrock (1945) considered limestone solution as the mode of origin for a sedimentary column of 15 feet. Irrespective of their origin the inorganic nature of the bottom sediments, red sand, clay and silt, arouses suspicion concerning a tundra or taiga pollen record.

While the southern Appalachians did not support local valley glaciers, tundra and taiga reached the Great Smoky Mountains (King and Stupka, 1950). In seeking landforms related to the piedmont swales one is drawn to the rather mysterious balds of the higher, south-facing summits. In view of the extensive ecological studies devoted to the balds and their origin (see review in Whittaker, 1956) it may be gratuitous to propose a new hypothesis. However, circumstantial evidence, i.e. their altitudinal distribution which parallels that of the blockfields and other periglacial features, and their well-known ability to resist tree invasion, places the balds in the category of possible full-glacial solifluction deposits.

In the west treeless flat valleys, the subalpine parks or mountain parks of the Rockies, occupy unglaciated mountains. Above 8300 feet in the Chiricahua and Pinaleno Mountains of Arizona, Barfoot Park, Rustlers Park and Hospital Flats are natural openings partly drained by semi-permanent streams in gentle valleys surrounded by coniferous forest. In their gentle slope, inorganic sediments and general aspect they resemble the Pennsylvania piedmont swales.

Daubenmire (1943) considered some of the parks as stages in primary hydro- or xeroseres. Others " . . . appear to be naturally inimical to tree growth; these are the true 'mountain parks' ". Ives (1942) felt that "Sub-

alpine wet meadows owe their origin to localized initial sluggishness of the valley drainage, leading to the formation of peat deposits." Beaver activity appeared responsible in part for their maintenance.

Evidently there is no common agreement on *why* trees are not rapidly invading the parks, glades, balds and swales (Daubenmire, 1943; Darlington, 1943; Whittaker, 1956). However, these natural openings appear to share various geomorphic features and may also share a common history. Should this prove the case they will vastly simplify the mapping of Pleistocene tundra and taiga zones in mid-latitudes.

Biotic and climatic zones during the full-glacial.—That European authors dare to offer vegetation maps of the glacial maximum (Frenzel and Troll, 1952; Büdel, 1949; Moreau, 1955) reflects their leadership in the field of Pleistocene paleoecology. Such an attempt in the New World would be premature. However, it may be possible to locate reference points to be incorporated eventually into such a map. From Iversen's experience with the Danish late-glacial (1954) we may anticipate that late Pleistocene vegetation zones were inhabited by combinations of species that no longer co-exist. This fact does not vitiate the assumption that climatic gradients induced biotic zonation during glacial as well as postglacial time.

Manley's paleoclimatic synopsis is a starting point for approaching the problem (1955, p. 260): "In light of the much more rapid gradient of summer temperature likely to prevail, it would thus be reasonable to suggest that inland in eastern North America trees might well be found within 80 miles of the ice margin, with well-developed forest not far behind." This would vary depending on exposure and influence of maritime air masses, i.e. a "cold wall" would send the tundra farther south in New Jersey than in the interior, just as the modern Labrador coastal tundra extends to lower latitude than the more continental tundra of central Ungava (Hare and Taylor, 1956). "On the above arguments it would be reasonable to expect that the forest would approach most closely to the ice along the flanks of the sheltered valleys in the western Allegheny Plateau" (Manley, 1955).

Most of Manley's deductions on the glacial maximum appear based on the absence of cirques in the southern Appalachians. This requires a sea level temperature of at least 56 degrees during the summer months and of 60 degrees during July. It means that the *coldest* plant formation we might expect in the Carolina piedmont and coastal plain would be boreal forest. The climate would be too mild for taiga (boreal woodland). Absence of Appalachian cirques also requires Manley to postulate a rather narrow periglacial tundra zone.

It is encouraging to find that both these paleoclimatic assumptions are born out by the full-glacial pollen record. The Marsh pollen profiles indicate very limited tree growth within 54 miles of the Wisconsin drift border. Studies of the Carolina Bay sediments (Frey, 1951, 1953, 1955a and 1955b) show that a boreal pine-spruce forest, and not taiga, occupied the coastal plain.

The Carolina Bay sediments unquestionably constitute our richest record of Pleistocene microfossils and environmental change in unglaciated eastern North America. The coldest interval is Frey's zone M2. Jack pine predomi-

nates; there are small amounts of spruce, composites, grasses and little else. This combination of species indicates an ecological analogue to the sand plain section of the Canadian boreal forest (Raup, 1946). Near the top of the profile (Frey, 1953) in late-glacial and early postglacial time oak and hickory appear with scattered beech, hemlock and sweet gum. A radiocarbon date (C-474, 10,224 ± 510) from this level corresponds roughly to the Valders readvance and shows that at this early date the Carolinas had warmed sufficiently to support temperate mesophytic rather than boreal forest.

Exact age of the M2 (boreal forest maximum) silt is unknown; however, it is bracketed by two radiocarbon dates, C-474 of 10,000 years and W-197 of greater than 38,000 years (Frey, 1955b). Almost surely M2 represents the last Wisconsin maximum in North Carolina. Tentatively I correlate it with the Marsh F zones.

A peak in non-arboreal pollen occurs in Frey's zone N. As he points out (1953, p. 307) the tree pollen shows a predominance of temperate species and there is no reason to claim that the NAP represents a taiga. I would attribute the peaks in alder and NAP and the high frequency of broken pine grains

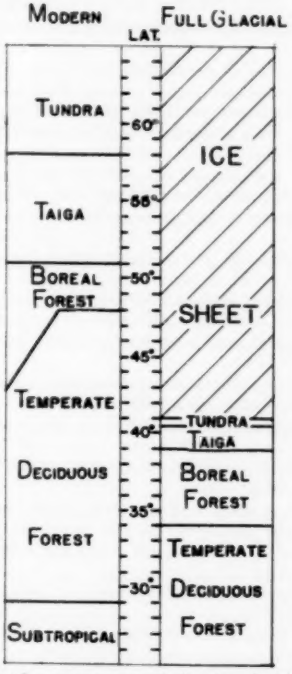

Fig. 17. Modern vegetation zones east of the Appalachians north to Baffin Island compared with the full-glacial displacement. The latter is based on pollen diagrams from Pennsylvania, North Carolina and on the absence of Appalachian cirques.

(Frey 1953, p. 303) to the same effect that accounts for their abundance in the upper Marsh sediments, i.e. deposition on a subaerial surface rather than in a lake.

Figure 17 represents a summary of the above three lines of evidence, i.e. Manley's deductions and the pollen diagrams from Pennsylvania and North Carolina. In the absence of stratigraphic or other proof of age I have rejected other diagrams from unglaciated North America. It is an obvious fact, but one not always made clear, that diagrams originating in unglaciated regions are not necessarily old. We now recognize that Darlington's Cranberry Glades profile is postglacial. Despite Potzger's claims (1945, 1952) there is no clear evidence that the southern New Jersey peat beds are any older. Perhaps it is worth repeating that pollen evidence of tundra can not be expected in peat or other highly organic sediments.

In figure 17 the present system of sea level vegetation zones east of the Appalachians and north to Quebec-Labrador is shown on the left. The full-glacial displacement, shown to the right, will appear conservative when compared to Peltier's map (1949) of a tundra climate reaching Virginia. It may seem radical to those who would locate elements of temperate forest close to the ice margin (Braun, 1955; Drury, 1956). I believe it is the interpretation required by present, fragmentary, paleoecological evidence.

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THE RELATION OF DEEP SEA SEDIMENTATION RATES TO VARIATIONS IN CLIMATE*

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ABSTRACT. Variations in sedimentation rates with time have been investigated in a mid-equatorial Atlantic core (A 180-74). Measurements of C^{14}/C^{12} ratios, percent coarse fraction (>74μ), and percent carbonate material at frequent depth intervals permit the calculation of absolute rates of sedimentation for each of the major contributors to the sediment. The rates of deposition of both the carbonate and clay fractions are shown to decrease abruptly from the high glacial rate approximately 11,000 years ago. The observed correlation of sedimentation rate with surface ocean temperature and extent of continental glaciation permits the dating of late Pleistocene events by reasonable extrapolations beyond the limits of radiocarbon dating. The beginning and end of the warm period preceding the last cold surface ocean water period are estimated at $\sim$ 150,000 and $\sim$ 70,000 years respectively.

INTRODUCTION

Deep sea sediment can generally be considered as composed of a calcium carbonate fraction and an aluminosilicate fraction, except locally where biogenic silica is important. (In this paper the aluminosilicate fraction is referred to as clay for convenience.) The factors affecting the rate of sedimentation of each of these components have been variously interpreted.

Arrhenius, et al. (1951) dated some cores from the Eastern Equatorial Pacific by assuming a constant rate of clay accumulation on the deep sea floor. The rate of clay accumulation was assumed to be directly related to the rate of titanium accumulation, which was determined using radiocarbon dates obtained in the more recent parts of a selected core from that area. This method is independent of variations in the abundance of other components in the core provided they are not titaniferous (Goldberg, 1954; Correns, 1954). Kullenberg (1953) has debated the premise that the clay sedimentation rate is constant. He assumes that the material being deposited from the water comes as particles from the land, and that the supply may have varied with climatic conditions. Arrhenius (1954), however, believes that the mechanism of accumulation of much of the clay in deep sea sediments is probably the coagulation of subcrystalline particles, and hence not subject to Kullenberg's calculations. Emiliani (1955) pointed out that the rate of sedimentation in certain Atlantic Ocean cores appeared to be lower during post-Wisconsin time than during the glacial period itself.

Wiseman (1954), postulating a constant rate of sedimentation for the clay fraction in the Equatorial Atlantic, suggests that the variation in the amount of carbonate relative to clay is the result of the varying productivity of calcium-carbonate-depositing organisms with varying ocean surface temperatures. For a pilot core from the Equatorial Atlantic (which he sampled to a depth equivalent to about 10,000 years), he found a decrease in the percent of calcium carbonate with depth which he related to the approach with depth to deposits of the last glacial period.

Urry (1948), using the ionium method of dating cores, concluded that in several cores the rates of sedimentation for all components had remained

* Lamont Geological Observatory Contribution no. 294.

sensibly constant with time. He particularly noted that the glacial periods did not accelerate or decelerate these rates. Volchok and Kulp (1957), using the same method of dating, found in several cores a greater rate of total sedimentation in post-glacial time than during the last glacial period.

Several workers believe that accumulation rates of one or more components of deep sea sediments have remained constant in some areas of the ocean and on this assumption have made estimates of the dates of Pleistocene events (Wiseman, 1954; Arrhenius, 1954). To test this assumption, we have made a detailed study of the nature of the variation of both clay and carbonate deposition rates in the Mid-equatorial Atlantic.

Although the results of this study are not directly applicable to the whole ocean, they do suggest some of the factors that may produce variations in other areas.

METHODS AND RESULTS

Selection of core.—Core A 180-74, selected for this study, was collected by Heezen during the 1952 cruise of the R. V. *Atlantis*. The core, one of a suite of four (the others being A 180-72, A 180-73 and A 180-76) taken on the eastern and western flanks of the mid-Atlantic ridge, appears to be ideal for this type of study, for there is no evidence of local turbidity currents, erosion, or solution of carbonate. Thus Ericson and Wollin (1956) show that the two cores taken on the east side of the ridge correlate in every respect with those from the west side, and hence that only particle by particle deposition took place.

Micropaleontological studies by Ericson and Wollin (1956) on this suite of cores reveal that the relative abundances of temperature-sensitive species of foraminifera vary with depth in the cores, indicating changes of surface ocean temperatures with time. Within the range of the carbon-14 dates obtained for these cores, the microfossils in each indicate a single major change in surface water conditions, *viz.*, a change (at a depth of approximately 20 cm for core A 180-74) from older relatively cold to more recent relatively warm surface water conditions. This change is thought to mark the end of the Wisconsin glaciation (Ericson et al., 1956). The existence and depth of this change have been confirmed by oxygen isotope temperatures obtained by Emiliani (1955) on one of the other cores (A 180-73) of this suite.

One of the primary aims of this study was to determine whether any variation in the rate of either carbonate or clay deposition was associated with this observed change in surface water conditions.

Measurements and calculation of rates.—Radiocarbon determinations were made on bulk carbonate at various depths in the core. Using the C^{14}/C^{12} ratio of carbon dioxide obtained from present-day surface ocean water as a control (Broecker, et al., 1958), ages were computed for the sampled segments. The radiocarbon results obtained are listed in column 3 of table 1 and are plotted in figure 1 as a function of depth in the wet core, together with Ericson's climate curve for comparison. Estimates of the rates of total sedimentation can then be made for successive intervals down the core. The sedimentation rates in centimeters per 1000 years have been computed for the

interval between each pair of radiocarbon dates by dividing the age difference by the depth difference (column 6, table 1).

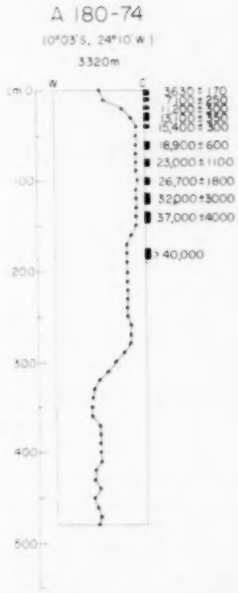

Fig. 1. Climatic curve obtained by Ericson and Wollin (1956) on the basis of relative abundances of temperature dependent foraminiferal tests for core A 180-74. The radiocarbon dates obtained by Broecker, Kulp, and Tucek (1956) and Broecker and Kulp (1957) for individual segments of the core are given in years before present.

In order to compare the rates of accumulation of materials of differing composition, porosity, and compaction all over the deep seas, it is more meaningful to express these rates in terms of the mass of material rather than the thickness, i.e., in terms of gm/cm^2/1000 yrs rather than cm/1000 yrs. The average bulk density of the dry core material of A 180-74 was determined to be 0.85 gm/cm^3. The density variation over the depth range of interest is less than 10 percent and hence can be neglected in the calculations which follow. Measurements of both length and diameter of the pieces of core before and after drying indicate a fifteen percent shrinkage in volume on dessication. Hence the bulk density of water-free material in the wet core must have been approximately 0.74 gm/cm^3. An additional correction of 5 percent in the density is made for the salt content of the dry material, determined by leaching a weighed portion of dry core material with a known amount of distilled water, followed by standard chlorinity titration. The salt corrected average bulk density for core A 180-74 is then approximately 0.70 gm/cm^3.

No sedimentation rate was computed for the top 2.5 cm of the core for two reasons. First, as much as five centimeters of core material may be lost

TABLE 1
Sedimentation Rate Data for Core A180-74

Sample Depth	Mid-Point	Age	Δ Depth	Δ Age	Total Sed. Rate cm / 10^3yrs	Total Sed. Rate gm / cm^2 10^3yrs	% Clay* Frac.	$CaCO_3$ Sed. Rate gm / cm^2 10^3yrs	"Clay" Sed. Rate gm / cm^2 10^3yrs	Percent Coarse Fraction (>74μ)	Coarse Frac. Sed. Rate gm / cm^2 10^3yrs
1	2	3	4	5	6	7	8	9	10	11	12
			>2.5	3630 ± 170							
0-5	2.5	3630 ± 170									
			8.0	3470 ± 300	2.3 ± .2	1.6 ± .2	11.0	1.4 ± .2	.18 ± .02	31	.43
9-12	10.5	7100 ± 250									
			9.0	4100 ± 390	2.2 ± .2	1.5 ± .2	16.6	1.3 ± .2	.25 ± .03	28	.36
18-21	19.5	11,200 ± 300									
			10.0	1900 ± 460	5.3 ± 1.3	3.7 ± .9	26.4	2.8 ± .7	.94 ± .24	20	.56
28-31	29.5	13,100 ± 350									
			10.0	2300 ± 460	4.3 ± .9	3.0 ± .6	29.6	2.1 ± .5	.89 ± .15	14	.29
38-41	39.5	15,400 ± 300									
			21.0	3500 ± 670	6.0 ± 1.1	4.2 ± .8	17.6	3.5 ± .7	.74 ± .14	19	.67
57-64	60.5	18,900 ± 600									
			19.5	4100 ± 1250	4.8 ± 1.5	3.4 ± 1.0	23.2	2.6 ± .8	.79 ± .21	21	.55
77-80	80.0	23,000 ± 1100									
			20.0	3700 ± 2000	5.4 ± 2.9	3.8 ± 2.0	19.0	3.1 ± 1.6	.72 ± .38	25	.78
97-103	100.0	26,700 ± 1800									
			19.5	5600 ± 3500	3.5 ± 2.2	2.5 ± 1.5	13.8	2.2 ± 1.3	.34 ± .20	31	.70
114-125	119.5	32,000 ± 3000									
			21.0	~5000	~4	~3	13.0	~2.5	~.4	27	~.7
135-146	140.5	37,000 ± 4000									
			41.0				15.0			31	
175-186	180.5	>40,000									

* salt-free dry material

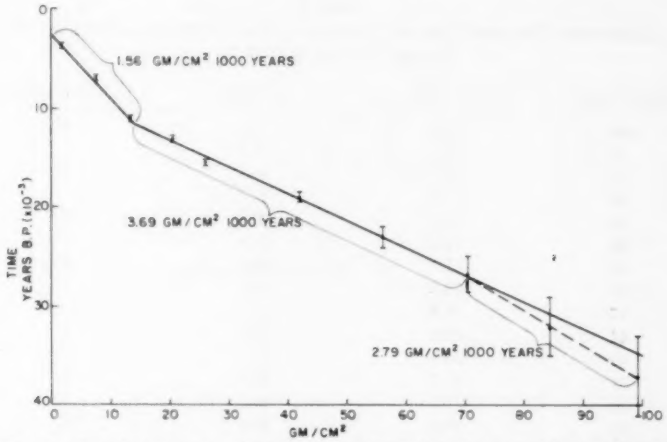

Fig. 2. Cumulative curve of total weight of salt-free dry material as a function of time in core A 180-74. The slopes of the straight line segments fit to the points are directly related to sedimentation rate.

during collection, and second, the top few centimeters of the material obtained are distorted to varying degrees during extrusion of the sediment from the coring tube. From visual observation, it appears that the effect of distortion in core A 180-74 probably has been quite small for the interval between the first and second samples and negligible for the intervals farther down the core. The sedimentation rate for each segment, expressed in gm/cm²/1000 yrs is given in column 7 of table 1.

In order to assess the carbonate and clay contribution to the total sedimentation rate, carbonate analyses were made on the dry core material at intervals of 5 cm, by the titration method described by Turekian (1956). After correction for the salt content of the dry material, the clay concentration is determined by subtraction. These results, together with the coarse-fraction data (greater-than-74 micron) from Ericson and Wollin (1956), are given in table 2.

The rates of carbonate and clay deposition for the interval between pairs of radiocarbon analyses are computed from the total sedimentation rate and the average carbonate and clay percentages (column 8, table 1) in the given interval. These results are listed in columns 9 and 10 respectively in table 1. The rate of coarse fraction deposition is given in column 12, table 1.

Errors.—The errors in the estimates of sedimentation rate are largely due to the uncertainty in the radiocarbon measurements. The error in the difference in age between the mid-points of adjacent samples is obtained from the square root of the sum of the squares of the errors on the individual ages. For the upper two intervals the uncertainty in the percent clay concentration contributes to the error. It should be noted that the sedimentation rates for the adjacent intervals are not independent; a change in the common age would

TABLE 2

Carbonate, Clay and Coarse Fraction (>74μ) Data as a Function of Depth in Core A 180-74

Depth (cm)	% $CaCO_3$	% "Clay"	%>74 Fraction*
top	93.6	6.4	30
6	89.7	10.3	
11	85.7	14.3	26
16	82.4	17.6	
21	79.7	20.3	23
26	72.5	27.5	
31	61.9	38.1	10
33	70.3	29.7	
37	79.4	20.6	
41	82.0	18.0	17
46	83.9	16.1	
51	81.2	18.8	15
56	84.3	15.7	
61	79.1	20.9	21
66	80.4	19.6	
71	72.5	27.5	16
76	76.4	23.6	
81	78.7	21.3	18
86	82.1	17.9	
91	81.5	18.5	28
96	81.1	18.9	
101	79.3	20.7	22
106	85.7	14.3	
111	88.3	11.7	32
131	87.4	12.6	22
151	85.3	14.7	28
171	82.8	17.2	24
191	88.9	11.1	30

* Ericson and Wollin (1956)

lower one value and raise the other. One way to avoid this is to make a cumulative plot of material deposited versus time. The slope of the curve that best fits the points gives the sedimentation rate. Such plots are shown for total material, for carbonate, and for clay in figures 2, 3, and 4 respectively.

DISCUSSION

Rate changes and climate.—From the data it is clear that a major change in sedimentation rate occurred in the equatorial Atlantic close to 11,000 years ago, the more recent rate being considerably less than the previous rate. The rate of clay deposition decreased by a factor of 3.7 and the rate of carbonate deposition by a factor of 2.1. The change takes place at the same depth in the core as the surface water temperature change. Although the data are limited, it appears that the sedimentation rate for both carbonate and clay has been nearly constant since the temperature change. Extrapolation of this rate suggests that 6 cm of material are missing from the top of the core. Post-deposi-

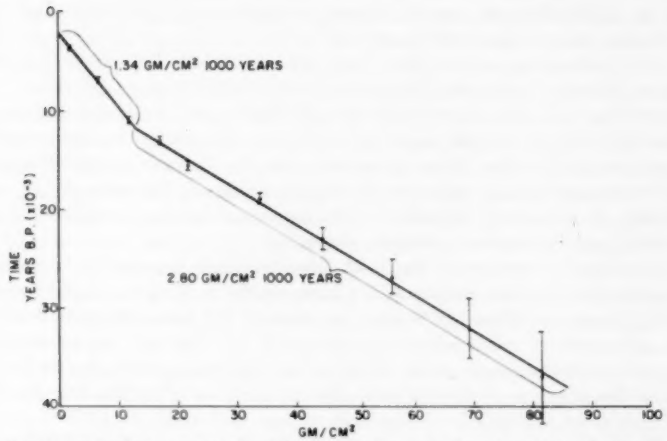

Fig. 3. Cumulative curve of weight of carbonate as a function of time before present in core A 180-74.

tional mixing of the sediment by bottom dwelling organisms may also explain part of this discrepancy. However, the presence of observable stratification in the core (Ericson and Wollin, 1956) minimizes the importance of this latter effect. Between 25,000 and 11,000 years ago adequate data indicate a uniformly high rate. For the period prior to 25,000 years ago the rate for clay deposition appears to have been intermediate between the high rate just preceding 11,000 years and the low rate just following it. Although there is a suggestion

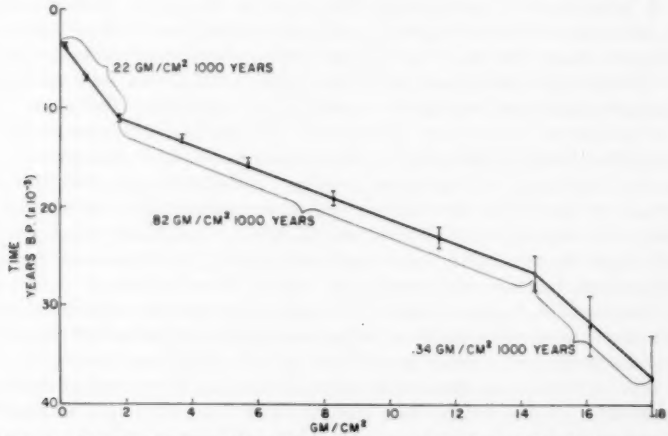

Fig. 4. Cumulative curve of weight of "clay" as a function of time before present in core A 180-74.

that the carbonate rate was also somewhat lower prior to 25,000 years ago, the change is not statistically significant.

It is interesting to note how well the rate of clay deposition correlates with the extent of continental glaciation. Prior to 26,000 years ago the ice was north of the Canadian border (Flint and Rubin, 1955). During the period between 25,000 and 11,000 years ago, the ice oscillated between the Canadian boundary and the Ohio River and finally after its final advance 11,400 years ago, it retreated rapidly into Canada and disappeared. The rate of clay (and probably of carbonate) deposition in the equatorial Atlantic seems to be highest during periods of most extensive glaciation.

Possible mechanisms.—The carbonate material consists of two major components, foraminiferal tests which make up the bulk of the coarse fraction ($>74\mu$) and coccoliths which make up most of the fine carbonate material. The agreement of radiocarbon measurements on both the coarse and fine fractions from a segment of one of the other equatorial cores (A 180-76) indicates the absence of appreciable detrital carbonate in the fine fraction (Ericson, et al., 1956).

From the size analysis data (table 2), it can be shown that 11,000 years ago the rate of coccolith (fine fraction $CaCO_3$) sedimentation decreased by a factor of 2.3. It is possible that the rate of sedimentation of foraminifera also decreased by as much as a factor of 1.6 at this time, but the spread in the data is great enough to make the reality of this change dubious. The problem is then to explain the decrease in these two components of the carbonate fraction. If it is assumed that the ocean is now saturated with calcium carbonate (Sverdrup, et al., 1942) and has been throughout the Pleistocene, two possibilities exist: 1) a redistribution of calcium carbonate deposition within the oceanic province, or 2) a general decrease in carbonate deposition rate, due to a decrease of calcium carbonate supply to the oceans from the continents.

A redistribution could result either from a change in internal circulation, as appears to have happened in the east equatorial Pacific (Arrhenius, 1952), or from changes in available shallow-water environments due to glacially controlled changes in sea level. A change in circulation would probably produce a change in nutrient supply, which would affect the productivity of the planktonic organisms. Alternatively changes in environments would conceivably involve a changing balance between deep and shallow water organisms competing for a common calcium carbonate supply. On the other hand, a net oceanwide decrease in carbonate sedimentation at the end of glacial time caused by decreased rate of calcium carbonate supply, might result from the decrease of erosional and runoff rates that must have accompanied the onset of the "interglacial" climate of recent times.

The coccolith accumulation rate is apparently more sensitive to changes in climate than the foraminiferal accumulation rate undoubtedly because the coccolithophores are autotrophic whereas the foraminifera are not.

The problem of explaining changes in the rate of clay accumulation is considerably complicated by the present uncertainties about the relative importance of mechanical transport versus chemical control in the formation of eupelagic clays. Mechanical transport to the oceans from the continents is by

wind or by streams, both agents whose effectiveness depends largely on climate. As to wind transport, loess accumulated fastest in Europe and North America (Flint, 1957) when the Wisconsin glacier was near its maximum. If loess accumulation is a measure of world-wide wind activity, then more material was supplied to the oceans by the wind during glacial time. Stream transport likewise may have been more important during times of expanded glaciers which are times of increased runoff and increased erosional rates. Also glacial lowering of sea level would reduce the area of the continental shelves, which act as sediment traps. As shown by Ewing, et al., (1958), glacial times are characterized by greater turbidity current activity; whereas the clay and silt load of the Mississippi River is presently trapped on the deltaic plain, during the glacial period it was deposited on the abyssal plain of the Gulf of Mexico. The transition is dated by radiocarbon at close to 11,000 years ago.

It is necessary to consider the possibility that some part of the deep sea clay may be of authigenic origin rather than detrital, as suggested by Arrhenius (1954) in defending against Kullenberg (1953) the idea of a constant rate of sedimentation of clay in the East Equatorial Pacific. The same sources that might supply varying amounts of detritus to the deep ocean should be expected to supply varying amounts of dissolved material as well. If clay minerals in some parts of the ocean are precipitated from solution as implied by Arrhenius (1954), then an increase in clay deposition might be expected where the concentration of dissolved material increases. The sea, however, is greatly depleted in silica and alumina. A two- or three-fold increase of dissolved material would still not satisfy the conditions for the chemical precipitation of silica in the sea (Okamoto, et al., 1957).

Possibly, however, silica and some alumina are removed from surface waters of the deep sea by diatoms. Variations in diatom abundance in surface waters would be a function primarily of nutrient supply. Some or all of the tests reaching the ocean floor may be transformed into poorly crystalline clay minerals during diagenesis by absorbing aluminum and magnesium ions from the sea in a manner suggested by Iler (1955) and Krauskopf (1956). The poor crystallinity of recent clays of the Atlantic has been demonstrated by Murray and Sayyab (1955).

In summary we must consider which of the mechanisms proposed above for the individual components may be simultaneously operative. Carbonate and clay sedimentation rates may vary sympathetically if both the nutrient and detrital supply were similarly affected by change in climate. If nutrient supply is increased by upwelling of deep waters, detrital material does not necessarily increase. Coastal waters, on the other hand, are rich in both detrital material and dissolved nutrients. Perhaps a major change in oceanic circulation with climate would permit a similar distribution farther out at sea. If, on the other hand, both the clay and the carbonate of core A 180-74 have their ultimate origin in organic life, the similar behavior of the two may be explained simply by the influence of a nutrient supply varying with climate. Perhaps, as a third possibility, the increase in both compoents is related directly to the greater supply of dissolved carbonate, nutrients, and detritus

which may have reached the deep sea during glacial times. It is obvious that only with more extensive studies of the sediments and circulation patterns of the Atlantic Ocean can a preference for one of the above mechanisms be reasonably defended.

On the other hand, the two hypotheses about rates of sedimentation mentioned by previous writers must be modified in the light of our data. First, Wiseman's (1954) correlation of high carbonate sedimentation rate and warm temperature in an Equatorial Atlantic core has not been borne out in our study. The greatest rate of carbonate accumulation (in the form of coccolith and foraminiferal ooze) is equated, according to our data, with the time of coldest surface water in the Equatorial Atlantic.

Secondly, Volchok and Kulp (1957) invoke a large sudden supply of detrital material at the end of the last glaciation to explain the gross increase in sedimentation rate since the last glacial period suggested by their radium data. From our results it appears that this sudden increase in the clay was of a short duration and had no significant influence on the sedimentation rate of post-glacial times.

APPLICATION TO PLEISTOCENE CHRONOLOGY

Several investigators (Arrhenius, et al., 1951; Suess, 1956; Emiliani, 1955; Wiseman, 1954) have attempted to construct chronologies of Pleistocene events from deep sea cores. Some of these early methods assumed either a constant rate of sedimentation of one component of the core sediment or an average rate of sedimentation for total core material throughout the Pleistocene. The "average" or "constant" rates were determined generally from a few radiocarbon measurements in the upper portion of the core and then assumed to apply for the rest of the core. Emiliani (1955) attempted to take into account the variation of rates between interglacial and glacial times. The limited amount of C^{14} data available at the time, however, made it difficult to adequately evaluate this variation.

Since we have demonstrated above that in at least one area of the ocean, the rates of sedimentation of all the major components vary considerably with climate, it is dangerous to use "average" rates of sedimentation or assume a "constant" rate of sedimentation for any component in inferring dates of older Pleistocene events.

It is possible, however, to use the information about the effect of climate on rates of sedimentation as a means of constructing a reasonable chronology at least for the late Pleistocene. The climatic curve for core A 180-74 (fig. 1) shows only two distinct events within the length of sampled core, *viz.*, the beginning and the end of the most recent cold surface water period. However, size fraction data indicate that the beginning of the preceding warm period lies immediately beyond the range of the sampled core. The end of the last cold period lies within the range of radiocarbon dating and has been determined to be about 11,000 years B.P. (Ericson, et al., 1956). The beginning of the cold period lies beyond the radiocarbon dating limit and must be estimated by extrapolation. The commencement of the cold period may correlate with the beginning of the Würm (Emiliani, 1955) or may merely mark the close

of a warm period within what is normally considered to be the Würm (Ericson and Wollin, 1956).

The estimate of the date of this change from warm to cold surface water conditions (chosen as the midpoint of the foraminiferal temperature curve) depends on the assumptions made. If the average rate of 5.25 cm/1000 yrs obtained for the period between 11,000 and 26,000 years ago is used for the entire glacial period, an age of 64,400 years is obtained. If, on the other hand, the intermediate rate of 3.7 cm/1000 yrs obtained for the period prior to 26,000 years is assumed to apply back to the beginning of the cold water period, an age of 80,700 years is obtained. Evidence presented by Flint and Rubin (1955), Emiliani (1955) and by Olson and Broecker (1957) suggests that an interstadial climate persisted during the middle portion of the last glaciation. It is generally concluded that this period ended close to 26,000 years ago (Flint and Rubin, 1955). The beginning of the period has not been well established. Unfortunately, this event is not observed in the foraminiferal record in core A 180-74, and hence the depts of the beginning and end must be estimated from other evidence. If it is assumed that this period began 45,000 years ago and ended 26,000 years ago (Brandtner, personal communication) and that during this period the rate of sedimentation was 3.7 cm/1000 yrs, an age of about 70,000 years is obtained. This age is calculated on the assumption that the interstadial was preceded by a glacial stage during which the sedimentation rate was the same as the 5.25 cm/1000 yrs calculated for the more recent glacial stage.

An alternate method for defining the time interval of intermediate conditions involves the use of the coarse fraction data. Ericson et al. (1956) and Ericson and Wollin (1956) from paleontological data and Emiliani (1955) from oxygen isotope paleotemperature data have shown a close relationship between surface water temperature and the percentage of $>74\mu$ material. The data presented in this paper explains this correlation by demonstrating that the variation in rate of coarse fraction deposition with surface ocean temperature is much less than that of the fine fraction. The percent coarse fraction is hence lower during cold surface water periods when the dilution by clay and coccolith material is relatively higher than during the warm water periods when the dilution is much lower. The data for core A 180-74 may then be interpreted as suggesting that mild temperatures persisted between a depth of about 100 cm and 210 cm. Radiocarbon dates then give an age of approximately 26,000 years B.P. for the 100 cm depth. Using the interstadial rate of 3.7 cm/1000 years the date of the beginning of the interstadial period is estimated as 56,000 years. Using the glacial rate of 5.25 cm/1000 years for the remaining 90 cm of the cold water period, an age of 73,000 years is obtained for the beginning of the last cold water stage.

Of the three, this latter age may be the best estimate in light of the present knowledge of late Pleistocene events. The limits for this event, as determined by the calculation for the extreme cases, are 65,000 to 81,000 years. This estimate does not differ greatly from that given by Emiliani (1955).

An estimate of the age of the beginning of the warm period preceding the last cold period can also be made. Emiliani (1955) correlates it with the

TABLE 3

Core	Glacial period			Earlier warm period		Ratio of glacial to warm sediment thickness	Avg. glacial rate cm/1000 yr	Avg. warm rate cm/1000 yr	Duration of warm period (years)	Age of beginning of warm period (years) (before present)
	End	Beginning	Thickness	Beginning	Thickness					
A 180-72	20	235	215	350	115	1.87	3.6	1.7	98,000	168,000
A 180-73	20	225	205	360	135	1.52	3.4	1.65	82,000	152,000
A 180-74	20	300	280	>480	>180	<1.56	4.7	2.25	>80,000	>150,000
A 180-76	20	260	235	410	150	1.57	3.9	1.85	81,000	151,000

All numbers, unless otherwise indicated, are in centimeters. The basic data are taken from Ericson and Wollin (1956). The method of calculation is given in the text.

beginning of the Sangamon Interglacial, while Ericson and Wollin (1956) suggest that it may mark the beginning of a major interstadial during the Wisconsin. In this case it is advantageous to use information from the entire suite of Lamont Atlantic equatorial cores (i.e., A 180-72, 73, 74, 76), because the cold to warm transition is not explicitly recorded within the limits of the studied core, A 180-74. The coarse fraction data as discussed above indicate, however, that the transition is undoubtedly just beyond the limits of the core.

It is observed (table 3) that though rates of sedimentation do vary among the cores of the Atlantic equatorial region, nevertheless, a constant relationship between the rates of sedimentation during glacial and non-glacial periods exists in all cores (except perhaps in core A 180-72). It is not possible to assume a varying rate of sedimentation for each core during glacial times and identical rates during the warm periods. In this regard the apparent constancy at the top of the cores for the present warm period is probably a reflection of the differing amounts of core top material lost. To test this it would be necessary to determine the present rates for each core from multiple radiocarbon measurements above the 20 cm zone, as was done in core A 180-74.

If we assume that the depth of material deposited during the last glacial period in each core represents 60,000 years of time, an average rate of gross glacial deposition can be calculated for each core. From core A 180-74 the ratio of glacial to non-glacial rates can be calculated. Assuming this ratio to be constant for all the cores of this suite, as discussed above, rates of sedimentation for the warm period previous to the last glacial stage can be calculated. Dividing this into the length of sediment represented by this period in each core gives an estimate of the duration of the warm period. These are given in table 3 for all the cores of this region, together with the absolute age (before present) of the beginning of the warm period preceding the last glaciation. If only core A 180-73, A 180-74, and A 180-76 are used, then an average age for the beginning of this period is $\sim$150,000 years. Inclusion of A 180-72 into the average raises it to $\sim$155,000. This estimate is 50 percent greater than that given previously by Emiliani (1955).

CONCLUSIONS

1. The rates of sedimentation for the various components in an Equatorial Atlantic core, A 180-74, have been determined using chemical, size fraction, and radiocarbon data. Table 4 is a summary of our findings.

2. From the data of table 4 and the paleontological work of Ericson and Wollin (1956), it is clear that the rate of sedimentation in core A 180-74 is related to temperature variations at the surface of the ocean. It is also related to extent of continental glaciation.

3. Both the carbonate and clay fractions show an abrupt decrease in sedimentation rate at the end of Wisconsin glaciation. The glacial to post-glacial ratio of sedimentation rates is 3.7 for the clay fraction and 2.1 for the carbonate. No definite conclusions can be drawn as yet about the cause of these changes in sedimentation rate for either fraction.

4. Table 4 shows that any attempt at time extrapolation into the Pleistocene using deep sea cores must be approached with a great deal of caution.

We have not found constant rates of sedimentation for any component, nor do we consider the concept of an "average" rate of sedimentation valid for large-scale extrapolation into the Pleistocene.

5. Using extrapolation procedures that depend on combining reasonable rates of sedimentation for each event in the late Wisconsin, we arrive at estimates of the date for the beginning of the last cold period ranging between 65,000 years and 81,000 year B.P. An additional extrapolation, using warm climate sedimentation rates, gives a date of approximately 150,000 years for the beginning of the warm period preceding the last cold water stage.

TABLE 4

	Post-glacial	Glacial	Last Inter-stadial
Clay	0.22	0.82	~0.35
Total Carbonate	1.34	2.80	~2.8
Foraminifera (>74μ)	0.40	0.57	~0.7
Coccolith (<74μ)	0.94	2.23	~2.1

The rates of sedimentation are given in terms of gm/cm²/1000 yrs. The Post-Glacial —Glacial transition occurs at 11,000 years B.P. and the Glacial—Last Inter-stadial break occurs at 26,000 years B.P.

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PERMAFROST FEATURES NEAR THE WISCONSIN GLACIAL MARGIN IN ILLINOIS

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ABSTRACT. Involutions and other frozen ground features at four localities in northern Illinois are described. Three of them are situated a short distance beyond the limit of Wisconsin glaciation and are striking in that involutions involve humic bands in Farmdale silt. The available evidence indicates that arctic climate existed briefly adjacent to the ice front in northern Illinois and that sporadic permafrost existed in a relatively narrow marginal zone. We conclude, however, that the arctic and tundra zones did not exist in the area of southernmost extent of Wisconsin glaciers and that the southern limit of permafrost was significantly north of the southern limit of glaciation.

INTRODUCTION

Cryoturbation (or congeliturbation) phenomena resulting from the effects of repeated freezing and thawing in the active zone above the permafrost table have been the subject of intensive study in arctic regions during the past two decades. Early during this period Sharp (1942) recognized involutions in unconsolidated late Pleistocene deposits in Illinois and attributed them to the effects associated with permafrost. Sharp reviewed the pertinent literature and established, in convincing terms, the permafrost origin of the features he described. These involutions occur within water-laid sediments of mid- to late-Wisconsin age, well within the limits of Wisconsin glaciation.

Horberg (1949; 1953, p. 32-33) described contorted humus bands in the Farmdale silts and certain wedge-like projections of overlying deposits into the silts. In discussing their origin he considered as a possible mechanism the action of the permafrost zone. He concluded that the wedge-like projections were the result of ice wedges, but rejected permafrost as the major cause of the contorted bands. Concerning their origin he stated (Horberg, 1953, p. 32):

> Since the deformed zones directly underlie Wisconsin till, the obvious interpretation of origin is by ice shove, although in places periglacial frost action may also have been an important process.

During the summer of 1957 we had opportunity to study involutions displayed by humus bands in the Farmdale silts adjacent to but beyond the limit of Wisconsin glaciation (fig. 1). It is our purpose here to describe briefly some of these features and discuss their probable climatic implications. The localities discussed here were selected for special attention because of their geographic location and the presence of thin humic bands that made it possible to record them by photography (pls. 1 and 2).

During part of the field study of these features we were in company with Jean de Heinzelin and Paul R. Shaffer. We express our thanks to M. M. Leighton, George E. Ekblaw, and Paul R. Shaffer who have read and criticized the manuscript.

STRATIGRAPHIC POSITION OF INVOLUTIONS

The permafrost features studied during 1957 occur within the Farmdale silts of earliest Wisconsin age. The Farmdale Loess was named by Leighton (1948) and its stratigraphic position as the first increment of Wisconsin deposition on the surface of the Sangamon Soil was clearly established (Leighton

and Willman, 1950). Although stratigraphic sections were measured at each of the localities studied, one only is presented here to illustrate the sequence of related deposits.

Section measured in fresh cuts in north bluff of Illinois River Valley, SW¼ NE¼ sec. 31, T. 7 N., R. 6 E., Peoria County, Illinois

	Thickness (feet)
Pleistocene Series	
Peorian Loess	
12. Loess, massive, upper five feet leached and contains soil profile, lower part weakly calcareous, tan to gray, streaked with limonitic root tubules, with scattered nodules and locally indistinct lamination	12.0
Farmdale Silt	
11. Silt, massive to laminated, weakly calcareous in mid-part but noncalcareous above and below, light gray; locally limonitic nodules and tubules; distinct thin humic bands in upper part; humic bands show distinct involutions throughout exposure (note pl. 1)	4.5
Illinoian Till	
10. Sangamon Soil. Soil A and G zones, leached; top one foot of A zone dark gray, granular to massive, grading downward into massive gley, gray mottled with tan	5.0
9. Sangamon Soil. Till, gray, thoroughly interlaced with veinlets of hard limonite, brown, with splotches of brown stain, weathering to a "box work" surface, noncalcareous in the gray matrix but locally calcareous in iron-cemented veinlets; "ferretto zone"	2.0
8. Till, calcareous, unevenly oxidized, well jointed, gray-tan, gradational at top and bottom	3.0
7. Silt and some sand, thin zones cemented with $CaCO_3$, red, tan, and gray. Southward the silts pinch out and a cobble zone is at same stratigraphic position	2.0
6. Till, oxidized, calcareous, jointed, brown	3.0
5. Sand and gravel in discontinuous lenses, locally cemented, brown	3.0
4. Till, calcareous, pebbly, blue-gray, well jointed throughout with oxidized rinds on joints	18.0
Kansan Till	
3. Yarmouth Soil. Soil, truncated to the B_2 horizon at top, leached, clayey, dark brown, locally spots of secondary carbonate; siliceous pebbles and cobbles present throughout; gradational with calcareous till at base and sharp contact at top except locally where blocks of this soil are incorporated in overlying till	4.5
2. Till, calcareous, gray with mottled patches of brown in upper part, jointed throughout with oxidized rinds along joints. Lenticular and irregular masses of gray to gray-tan silt and very fine sand with contorted and disorderly oriented bedding occur discontinuously in the middle part; although no fossils were found at this locality, these silts strongly resemble fossiliferous silts in a similar stratigraphic position in nearby exposures in SW¼ NE¼ sec. 11, T. 6 N., R. 5 E., Fulton County	18.0
1. Sand, fine- to medium-grained, calcareous, brown with locally dark gray streaks at top, upper contact irregular. Base of section in diversion ditch at base of exposed face	4.0
Total thickness exposed	79.0

Inasmuch as three of the four localities discussed herein lie near but outside the limit of Wisconsin glacial advance (fig. 1), the Farmdale silt at these localities was not overridden by any glacier and therefore the contortion of

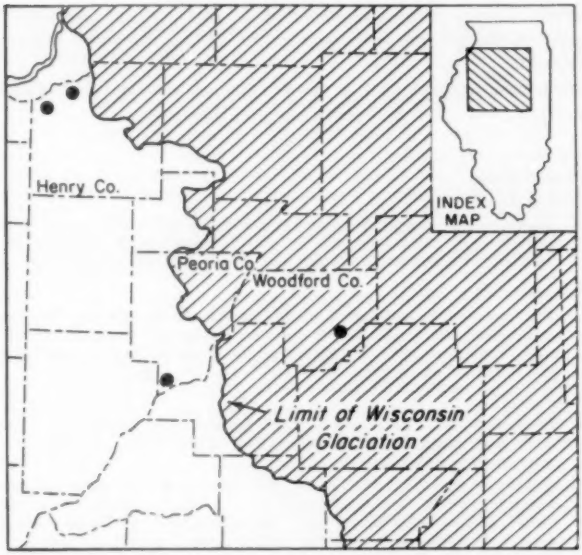


Fig. 1. Map of a part of northwestern Illinois showing location of the exposures described.

the humic bands cannot be attributed to ice shove. The stratigraphic and geographic setting indicates that the Farmdale silt was near the surface when the glacier front was just a few miles away.

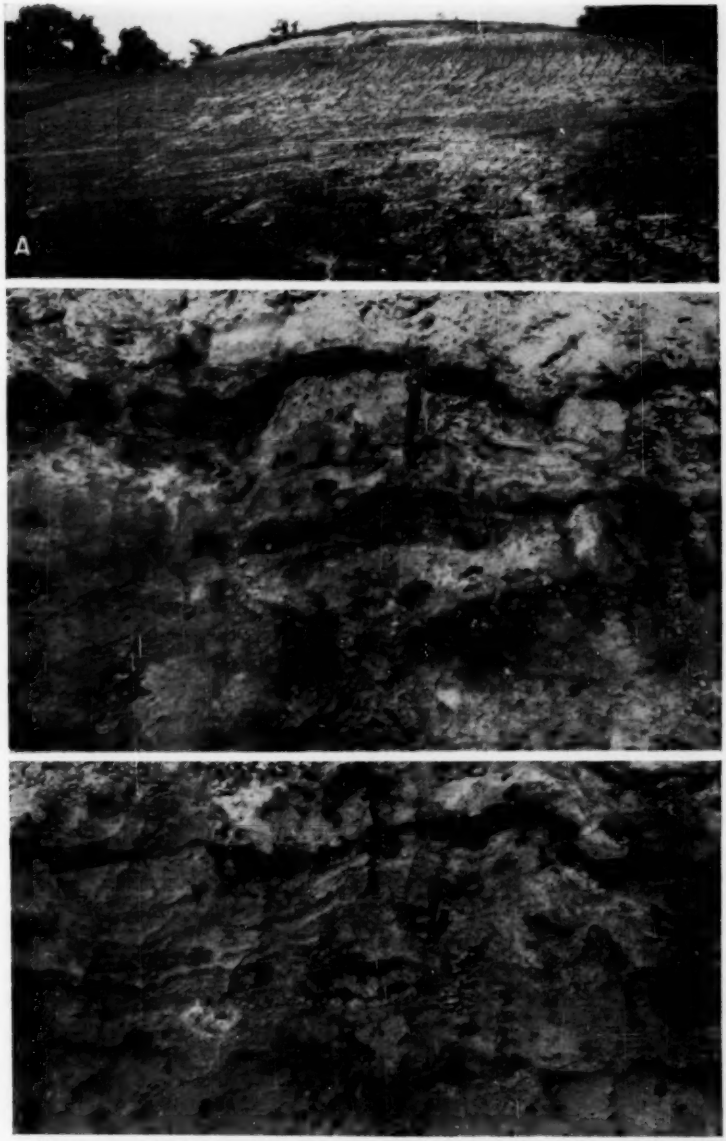
DESCRIPTION OF FROST FEATURES

At the exposure in Peoria County, the southernmost locality discussed (fig. 1, pl. 1), the involutions are less intense than those displayed in Henry County (pl. 2, B and C) but are more clearly shown by the thin zones of humic staining. Although the involutions were present throughout 50 yards of continuous exposure they were uniformly of a mild type without "overfolding" or the isolation of "pods."

In contrast, in the exposures studied in Henry County (pl. 2, B and C) the involutions display "overfolding," "infolding," and isolation of "pods" surrounded by humic material, but in the photographs this higher degree of deformation is not so clearly evident because of the generally greater thickness of the humic zones.

Still different conditions are exhibited in Woodford County (pl. 2, A). Thin Farmdale silt above Sangamon Soil exposed in road cuts contains humic bands but does not contain involutions, even though the locality is more than 30 miles within the limit of Wisconsin glaciation. The Farmdale silt is overlain by an upward succession of thin Iowan Loess, a thin sand zone, and calcareous Shelbyville till. Both the loess-sand contact and the sand zone at the base of the till are contorted, and wedge-shaped cracks in the Iowan Loess are filled with the overlying sand.

PLATE 1



Exposures of Pleistocene Deposits Along West Bluff of Illinois River Valley, SW¼ NE¼ sec. 31, T. 7 N., R. 6 E., Peoria County, Illinois.

A. Bluff line exposure, locality of measured section and involutions shown in B and C.

B. and C. Involutions in loess with peaty bands in the upper part of the Farmdale Loess: pocket knife gives scale.

INTERPRETATIONS

A divergence of opinion has existed for many years concerning the nature of the climate adjacent to the margins of continental glaciers that advanced into relatively southern latitudes. The European literature in general indicates the conclusion that arctic climates and permafrost existed rather extensively beyond the limits of the ice during glacial epochs. On the other hand, Frye and Leonard (1952) have presented evidence indicating relatively mild climatic conditions in close proximity to the glacier front during the maximum extent of both the Nebraskan and Kansan glaciations in their type region.

The cryoturbation features here described furnish evidence bearing on this problem in relation to the Wisconsin glaciers in Illinois. The lack of involutions in the banded Farmdale in the Woodford County exposure, coupled with the permafrost features in the thin overlying Iowan Loess, suggests that frozen ground existed just prior to the arrival at that locality of the first Wisconsin ice front (Shelbyville) but for such a short time that associated deformation did not extend as deep as the Farmdale. Furthermore, the involutions in the Peoria and Henry counties exposures are judged to extend upward into the basal (Iowan) part of the overlying Peorian Loess, and thus confirm the tentative dating as the time of maximum ice advance, and for a brief period only.

The lower part of the Peorian Loess is involved in the deformation, but its homogeneous character prevents definite recognition of involutions far above the base. Outside the area covered by Wisconsin glaciers, the Iowan Loess is so similar to the overlying loess that it generally cannot be differentiated, hence the use of the name Peorian for the undifferentiated Wisconsin Loess above the Farmdale. As 5 to 10 feet of Iowan Loess was deposited widely in this region before the Shelbyville ice arrived, the thickness of the zone affected by permafrost may be of that order. It is possible, however, that the Iowan Loess was abnormally thin at the localities in Henry and Peoria counties. Furthermore, frost features have been found at so few localities that permafrost may have been sporadic and formed only in places where topography and drainage were unusually favorable.

Correlation of these relatively rare frost features with the immediately adjacent Shelbyville moraine, one of the largest of the Wisconsin moraines, suggests that such structures may be found only near positions of moderately long stands of the ice front. If a narrow permafrost zone developed along the ice front, it seems probable the ice advanced rapidly enough to cover it, or retreated rapidly enough to abandon it, before significant deformation such as involutions could be formed. It appears that the most favorable places to find permafrost features in Illinois would be close to the larger moraines, such as the Bloomington and Marseilles. The only other occurrences of permafrost

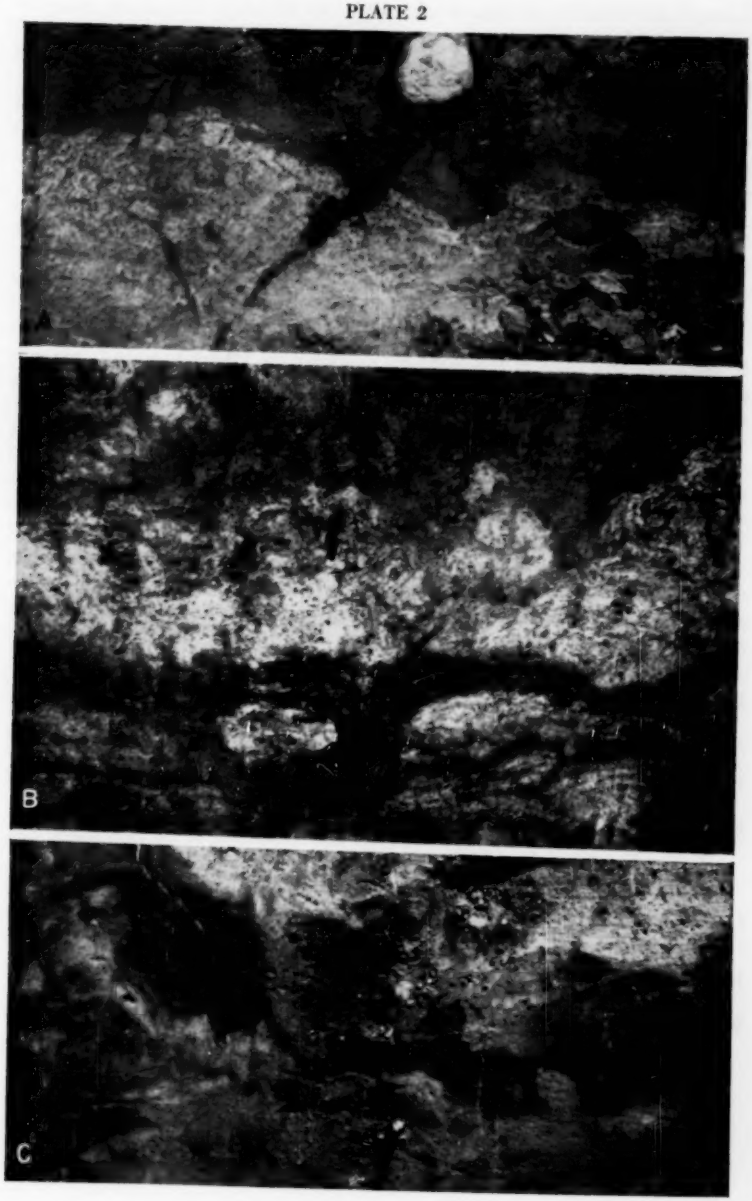
Frost-features in Loess. →

A. Ice wedge in loess below Shelbyville till, NE¼ sec. 9, T. 26 N., R. 1 E., Woodford County, Illinois.

B. Involutions at the top of the Farmdale Loess, cent. N. line NE¼ sec. 2, T. 17 N., R. 2 E., Henry County, Illinois.

C. Involutions at the top of the Farmdale Loess, SW cor. NW¼ sec. 21, T. 17 N., R. 1 E., Henry County, Illinois.

PLATE 2



features described from Illinois are related to the large Valparaiso moraine (Sharp, 1942).

Within the Wisconsin drift many layers of sand and silt interbedded with till are believed to result from repeated retreats and advances of the ice front. The preservation of these silts suggests that the ground may have been frozen when they were overridden, but the absence of deformation suggests that permafrost could not have existed in them for many years before burial by the glacier.

That an arctic climate could not have extended far from the ice is shown by the abundant snail fauna in the Peorian Loess in locations which could not have been more than 25 to 50 miles from the ice front during the major accumulation of the loess. In some locations proximity of the fossiliferous loess to the ice front may be less than 10 miles.

The fact that the intensity of involutions in the Peoria County exposure is markedly less than at the northern exposures in Henry County is at least suggestive. Cryoturbation features have not been observed in the many scores of exposures of this zone examined along Illinois Valley farther south.

Based on the foregoing data and the observations of Sharp (1942), permafrost existed for a relatively short time, and perhaps locally, adjacent to the margins of Wisconsin glaciers in northern Illinois, but became ineffective in central Illinois. It is judged that an arctic climate was not prevalent at the Wisconsin glacier border in its southernmost extent.

The inference seems clear that in this latitude the ice had advanced into a region so far from climatic conditions which favored its growth, that the arctic and tundra zones may have been effectively eliminated, rather than merely compressed.

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DISCUSSION

THE NOMENCLATURE OF EVAPORITE TEXTURES

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The recent paper and discussion about the terminology of evaporite textures (Greensmith, 1957, 1958; Stewart, 1958) raises what has long been a vexed question.

Petrology must surely be one of the few sciences in which descriptions are given in terms of what we interpret the origin to have been. Thus, using Dr. Greensmith's example, if we *think* a rock is igneous in origin we describe a certain texture as poikilitic, but if the same texture occurs in a rock which we *think* is metamorphic then we term it poikiloblastic. We must welcome Dr. Greensmith's plea for a non-committal terminology but let us not define new terms for a small and special group of rocks, evaporites, while continuing to use genetic terminology in the whole of the remaining field of igneous, metamorphic and sedimentary petrology. Dr. Greensmith's proposals amount to a new set of terms to be used for a particular rock of a particular origin, just as igneous terms are used for igneous rocks and metamorphic terms for metamorphic rocks, whereas I feel descriptions of a particular texture ought to be the same whatever the type of work.

As Niggli (1954, p. 235) says in proposing a comprehensive and rational terminology for the whole of petrology, "Crystalloblasts are formed no less frequently [in metamorphic rocks than] in the late products of magmatic consolidation, and limestones, too, as well as other sediments, often exhibit a crystalloblastic fabric. When developments call for similar descriptions and are of similar genesis there is no justification for giving them different names merely because they appear in a magmatic rock on one occasion and in a metamorphic rock or sediment on another. By so doing we would tend to hide the fact that very similar processes can be detected in all three types and that it is often very difficult to deduce the genesis of a rock as a whole from a study of its fabric".

Clearly the common feature of metamorphic rocks, many evaporites and limestones, and some granites is that they have recrystallised and our description of the resulting texture ought not to vary with the type of rock recrystallized.

Finally, in answer to Dr. Greensmith's point about the teaching and students' point of view, I think it is easiest and clearest to follow Niggli's (1954) scheme using the same terms for the same textures, whatever the environment. One does not then need to know the origin of a rock in order to be able to describe it!

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